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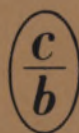
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to have taken place reasonably soon thereafter.*

In the article, I argue that pelagic lithogenesis should be classified as a separate type of sedimentary process constituting a portion of oceanic lithogenesis in the sense suggested by N. M. Strakov. Distinctive attributes and criteria for distinguishing pelagic lithogenesis are discussed and regions where the process occurs, open areas of large ocean basins, are delineated. The following stages in pelagic lithogenesis are distinguished: preliminary working and supply of sedimentary material, sedimentation, water-ocean floor contact, diagenesis, and catagenesis. It is shown that stages of sedimentation dominated by a biogenic mechanism and processes at the water-ocean floor contact are key stages in this process.

N. M. Strakov [23, 24] completed his theoretical treatise on sedimentary processes over the entire Earth by proposing a separate classification for a sedimentary oceanic type of lithogenesis. In the opinion of N. M. Strakov [25, p. 185], the known features of the sedimentary process in oceans "allows us to distinguish a special oceanic type of lithogenesis different from all types of lithogenesis associated with the continental block," i.e., different from glacial, arid, and humid types. In his opinion, there is a single sedimentary process at work in the ocean. This process is governed by the hydrodynamic regime of surface waters, and is combined with a process of volcanogenic-sedimentary lithogenesis which is the same on continents and in ocean basins.

The controversy building up around the concept of *oceanic lithogenesis* at that time was cut short by N. M. Strakov's death. The goal of this article is not to renew this controversy, although the questions raised in it have not been satisfactorily answered. Some of the questions are discussed below in the light of extensive new data on oceanic sedimentation appearing the last decade, including the author's own results summarized in a monograph [14] and in other publications. It is not possible to present the data itself here; therefore, I will discuss, in the order raised, the problems and explanations associated with the author's viewpoints as developed over more than 30 years of investigations of oceanic sedimentogenesis in conjunction with the staff of marine geologists at the P. P. Shirshov Institute of Oceanology, USSR Academy of Sciences, which was headed by P. L. Bezrukov for many years.

I have attempted here to develop a key of idea of N. M. Strakov's concerning specifics of lithogenesis in oceans as they relate to the huge dimensions of oceanic sedimentary basins. The question of the correctness and expedience of creating a new classification for a special *pelagic* type of lithogenesis will be discussed. This conception, in the opinion of the author and a number of his colleagues investigating sedimentation in the modern ocean, incorporates all of the important features found in N. M. Strakov's oceanic type of lithogenesis; at the same time, processes occurring at the margins of ocean basins (in the near-continent region, in our terminology [14]) remain outside this conception. We have thus succeeded in removing one of the contradictions to the idea of a single oceanic type of lithogenesis in which a typomorphic *pelagic* sedimentary process fully characteristic only of open (pelagic) regions of large ocean basins turned out to be combined with processes in near-continent regions similar to humid processes and, in individual regions, similar to the arid or glacial types of lithogenesis.

It should be emphasized that the author uses the term *pelagic* in the sense in which the term was used in marine geology at the time of the Challenger expedition [33]: to refer to processes (involving sediments and facies) of the open ocean possessing a specific set of properties and attributes [14]. We propose use of the term *hemipelagic* to refer to sedimentary processes in the central areas of deep seas and also in the near-continent regions which are similar in their mechanism to pelagic processes (i.e.,

processes taking place primarily via precipitation of sedimentary material from the water column-pelagic zone), since they differ sharply from typical pelagic processes in intensity, character of diagenesis, and a number of other attributes. We will discuss this distinction below in more detail when reviewing the stages of pelagic lithogenesis.

The author adheres to the N. M. Strakov's point of view, combining all stages of the sedimentary process, from mobilization of material and sedimentation to conversion of sediment to rock, in the concept of lithogenesis. I believe the other, narrower understanding of the term, which became current after the classical work of N. M. Strakov, to be unworkable.

In order to justify classifying pelagic lithogenesis as a separate type, it is necessary to demonstrate its specificity as a process of sedimentary rock formation at all stages, from mobilization and transport of the original sedimentary material to the accumulation of sediments and their conversion to rock during diagenesis and catagenesis. N. M. Strakov repeatedly emphasized that type of lithogenesis is a genetic, rather than geographic concept, and that criteria for distinguishing types should not simply coincide with climatic zones; they should rather coincide with intrinsic differences between the processes at all stages, as reflected in sedimentary rocks and their geochemistry and mineralogy (above all, where authigenesis is involved). We will attempt to show that pelagic lithogenesis as a process possesses all of the attributes needed to classify it as a separate type.

Distinctive Features of Pelagic Lithogenesis

The concept of *pelagic lithogenesis* was proposed in thesis form by P. L. Bezrukov, jointly with the author of the present article, approximately 20 years ago on the basis of then existing conceptions of the sedimentary process in the ocean. Subsequently, however, we, like the majority of lithologists working in the ocean, concentrated our attention on processes of sedimentogenesis, their zonation and facial heterogeneity; other stages of lithogenesis escaped our field of view. Instead of the term *lithogenesis*, we accordingly used the familiar concept of pelagic sedimentogenesis, although always keeping in mind the specifics associated with diagenesis (especially early diagenesis) of pelagic sediments as compared to nearcontinent sediments ([14], etc.).

Pelagic sediments were already classified in a special lithological-facial group contrasted to terrigenous sediments in the first classifications of oceanic sediments [33]. Their primary features, even according to newer points of view, were presented in a classical work [27]. The author considers pelagic sediments to be a special facial complex (pelagic megafacies) characteristic only of oceans.

The main distinguishing features of pelagic sediments are considered to be oxidation (red coloration), low content of organic matter, high contents of iron, manganese, and a number of microelements (recalculated for the abiogenic portion of the sediment), a specific complex of authigenic minerals (ferrismectite, phillipsite, various mineral forms of iron hydrates and manganese, especially in the form of nodules, barite, etc.), and an absence of terrigenous material, other than fine pelite. All types of pelagic sediments, except for pelagic (red) clays, are, for the most part, biogenic, composed of skeletal remains of pelagic planktonic microorganisms: carbonate (foraminifers, coccoliths, and pteropods) and siliceous (diatoms, radiolarians, and silicoflagellates). In addition to purely lithological features, criteria for the identification of pelagic facies include low rates of sedimentation and small absolute masses of all components, especially of terrigenous material and C_{org} , small absolute masses of microelements associated with them, an absence of fossilized remains of benthic macrofauna and the presence of bone detritus and teeth of pelagic ichthyofauna. Breaks in sedimentation are characteristic, along with the presence of numerous areas of nondeposition and even erosion on the modern ocean-floor surface at all depths; breaks in sections of pelagic sediments penetrated by deep-water drilling are characteristic.

We can see that the majority of these objectively observable (or measureable) attributes coincide with those put forward by N. M. Strakov as criteria for identification of oceanic lithogenesis. At the same time, they are all in contrast to the attributes of the precontinental oceanic sedimentary process.

Indeed, the precontinental megafacial region [14] is characterized by reducing conditions of early diagenesis, including sulfate reduction and formation of authigenic sulfides, high rates of terrigenous sedimentation (up to the point of "avalanches" [11]), high concentrations and absolute masses of organic matter, low concentrations of iron and, especially, manganese, with high values for their absolute masses. The earlydiagenetic mechanism of reducing remobilization and migration of manganese, typical of humid lithogenesis, is operative here. Associations of authigenic minerals "prohibited" in the pelagic, deeply oxidizing zone, occur here; such minerals include: iron sulfides, glauconite, and carbonates of iron and manganese. One can scarcely describe such profound differences as those between pelagic and near-continent megafacial regions to variations of a single type of oceanic lithogenesis, especially if the similarity (in fact, complete match) between attributes of near-continent oceanic and humid marine lithogenesis is taken into account.

The differences between pelagic and near-continent lithogenesis become even more profound when we turn from the features recorded in the sediments to the processes giving rise to them during the various stages of sedimentary rock formation distinguished in N. M. Strakov's theory of lithogenesis.

Stages of Pelagic Lithogenesis

When characterizing oceanic lithogenesis, N. M. Strakov voiced the opinion that the specifics of post-sedimentational processes in pelagic regions of the ocean resulting from different rates of sedimentation and oxidation do not allow us to use "the scheme of stages which was developed for the usual reduced sediments found in water bodies in humid climates or even the peripheries of the oceans themselves" [25, p. 185]. The latter point is especially important in this context, for it once again shows N. M. Strakov's understanding of the specifics of pelagic lithogenesis as compared to the near-continent ocean, an understanding which is certainly not limited to post-sedimentational processes but rather encompasses all stages of lithogenesis, including the supply of the original sedimentary material, its transport, differentiation, and fixation in sediments during the course of lithogenesis. The stage of lithogenesis itself is not lost in our classification; the boundaries between stages are clearly differentiated, although the stage of lithogenesis includes different content than in N. M. Strakov's scheme of lithogenesis.

Mobilization of sedimentary material, the first stage in the climatic types of lithogenesis on a continental block, is not directly expressed in pelagic lithogenesis. The major portion of the original sedimentary material is subjected to processes found in other types of lithogenesis, is supplied to the system of pelagic sedimentation from without, and is already transformed to a significant degree. Products of continental denudation (terrigenous material) and volcanism (volcanoclastic material and exhalation products) are the primary sources of this material. Moreover, exchange at the ocean-atmosphere interface, controlling the regime of O and CO₂ in the ocean, is important to the balance of the original sedimentary material where these components participate in sedimentation processes through physical-chemical and biological systems. Of the internal primary sources of sedimentary material in pelagic lithogenesis, halmyrolysis (submarine chemical weathering) and mechanical breakdown of the ocean-floor bed rock, accompanied by formation of edaphogenic clastic material (mainly under the influence of tectonic processes) plays a significant, although subordinate role [13, 14].

The terrigenous source is linked through climatic conditions associated with denudation of land to humid, arid, and glacial types of lithogenesis of the continental block; this is reflected in a latitudinal (climatic according to A. P. Lisitsyn) zonation in the supply of solid terrigenous material to the ocean [8-10, 14]. Thus, eolian dust, with its own characteristic mineralogical and structural features, is predominantly supplied from arid zones of the continents and is dispersed over the ocean by latitudinal atmospheric currents, thereby leaving its mark upon the mineral composition of the pelagic sediments in the corresponding latitudinal zones. The products of glacial lithogenesis, dispersed by icebergs, sea ice, and also in the form of eolian and aqueous suspensions, reach pelagic regions in the polar latitudes. Fluvial (transported by rivers) suspensions are supplied in large quantities from the humid zones of the continents.

In our opinion, this zonation in the supply of solid terrigenous material cannot serve as a sufficient basis for distinguishing various climatic types of lithogenesis in the ocean (especially in the pelagic region of interest to us here). In the first place, this is because climatic marking of terrigenous material is so weak against the background of more pronounced patterns due to other causes. Orography and geological structure (the rock resources) of distributive provinces on continents [7, 14, 25] and also the action of barriers and sedimentation traps on the path of the terrigenous from land to pelagic regions of the ocean are applicable here.

In the second place, pelagic lithogenesis is predominantly fed by dissolved components directly leached by biological and physical-chemical processes from the generalized "dynamic reservoir" of ocean (in the terminology of A. P. Lisitsyn). The dynamic reservoir functions as a system of material circulation controlled by hydrodynamic, biological, and physical-chemical processes taking place in the ocean and only indirectly associated with climatic conditions.

Pelagic lithogenesis develops only under conditions of low supply of terrigenous material consisting of the most finely dispersed fractions. Remoteness of regions of pelagic lithogenesis from the sources of terrigenous material sufficient to meet these conditions occurs only in large (tectonically "mature") ocean basins, where belts of near-continent, predominantly terrigenous sedimentation are located between the shores of the continents and the pelagic region. According to A. P. Lisitsyn [11], no less than 92% of the solid terrigenous material removed from the continents (the estimate more likely to be low rather than high) settles in this belt, belt which can reach widths of many hundreds and even thousands of kilometers and includes shelf, continental slope, aggradation bodies (fans and aprons), the foot of the continent, and near-continent aggradation plains [14].

The near-continent region not only acts as a trap for the majority of terrigenous material (including finely dispersed clayey material) but also acts as a filter separating dissolved components of fluviogenic terrigenous runoff from suspended matter and separating the finest suspension from the coarser-grained fractions, thus preparing the original terrigenous material which is supplied to the system of pelagic lithogenesis.

The role of eolian and glacial (including iceberg) material becomes significant and, in some cases, dominant, against the background of low supply of fluviogenic suspension to the pelagic region [8, 9, 14]. The supply of eolian material is realized via atmospheric currents and, of course, is not exposed to the action of the near-continent "filter." This material is subjected to mechanical differentiation in the atmosphere, as a result of which only the finest dust reaches extremely remote pelagic regions.

Transport of terrigenous material to the pelagic region by sea ice may be of some importance only in the Arctic Ocean. Iceberg dispersal, serving as the main supplier of terrigenous material to the southern near-polar zone of the pelagic region in the modern ocean, and, Pleistocene continental glaciation acting in the Northern hemisphere, play much greater roles. Material from iceberg dispersion, as is well known, is granulometrically unsorted and bears indications of glacial lithogenesis [8, 9].

Volcanogenic material from subaerial eruptions reaches pelagic regions of the oceans by two routes: through the atmosphere in the form of fine ash transported by air currents in a manner analogous to the transport of terrigenous eolian material and in the form of floating pumice dispersed by surface currents. The second route is especially important [14], providing not only broad distribution of pumice fragments in pelagic sediments, fragments often serving as nuclei for ferromanganese nodules, but also promoting formation of tephritic pelagic sediments and interbeds of vitroclastic tephra.

Submarine exhalations associated with ocean-floor spreading make their own contribution to the balance of dissolved components in the dynamic reservoir of the ocean, and thus to pelagic lithogenesis. The quantitative contribution of exhalational sources of different elements remains debatable, but it is important to note here that only the portion of components carried out by exhalations which does not precipitate near the sources and which is incorporated into pelagic circulation will become involved in pelagic lithogenesis (i.e., will become involved in processes of volcanogenic—sedimentary lithogenesis).

During the initial stage, pelagic lithogenesis is thus characterized by primary sources of sedimentary material (mobilization of material) and processes of sedimentogenesis separated in time and space; there is an immediate break in the source—sedimentation connection. Between these two stages, a colossal buffer in the form of the dynamic reservoir of the ocean and its circulation system is active.

Transport and transformation of the original sedimentary material in the water column of the ocean is specific to pelagic lithogenesis, but is more or less uniform over the entire ocean, except for the narrow belt of coastal shallow waters. The pelagic transport mechanism is realized by horizontal and vertical circulation of ocean waters, quasi-steady-state surface currents playing the main role in movement and redistribution of suspended matter; distribution and transformation of dissolved components is controlled primarily by vertical circulation through the biosystems of the pelagic zone. The biological mechanism of transformation of dissolved but, to a significant degree, suspended matter is the basis of pelagic sedimentogenesis. It is precisely here that we find the intrinsic difference between the pelagic type of lithogenesis and all other types of sedimentary processes, including humid marine and near-continent oceanic processes, where biogenic processes have important, although not definitive significance.

The pelagic biogenic process, a process in which planktonic microorganisms (phyto- and zoo-plankton and also bacteria) play the main role, completes the complex and many-faceted work of transforming matter and energy. The main geological consequence of this process is the generation of sedimentary particles capable of settling to the bottom through a multikilometer stratum of moving ocean water under the influence of gravitation. Two functions of planktonic organisms manifest here. On the one hand, certain of their groups create solid, poorly dissolving skeletal elements or shells (calcareous, siliceous, more rarely phosphatic) by converting soluble forms of the appropriate chemical elements into solid (suspended) forms. Besides the main sediment-forming components, mineral particles generated during the life activity of planktonic organisms include (in different forms of binding to a carrier) a specific set of microelements important for sediment formation, including heavy metals; planktonic organisms facilitate their extraction from the dilute solution of ocean water.

On the other hand, as is now certainly established, organisms of zooplankton filter finely dispersed suspended particles of both biogenic (coccoliths, diatoms, etc.) and terrigenous (clayey and detrital minerals) origin during feeding, ejecting the "unedible" portion in the form of molded balls covered with an organic film (pellets). As demonstrated in recent years, the biofiltrational (pellet) mechanism for settling out fine suspension occupies a leading role in the system of pelagic sedimentation [5, 12]. It promotes comparatively rapid settling of organic matter incorporated into the pellets, as well as mineral particles, to the bottom.

The intensity of any biological processes in the pelagic zone depends upon primary production of phytoplankton as a main food resource. The distribution of primary production in the ocean, as is well known, is controlled by the supply of rich nutrient substances (phosphates, nitrates, and silica) from subsurface or deep waters to the upper layer of photosynthesis as a result of divergences, upwellings, or mixings under conditions of seasonal convection in zones of geological fronts; these nutrient substances control and limit the development of phytoplankton. The field of primary production and, consequently, biological productivity of the ocean in general, possesses clearly expressed features of circumcontinental and latitudinal zonation. Both zonations are hydrodynamic in nature and are due to vertical circulation in the upper layers of the oceanic water column leading to "fertilization" of the layers; they are only indirectly associated with climatic factors. The latitudinal zonation of bioproductivity (like the water circulation) is clearly expressed in the pelagic regions of three oceans, whereas the circumcontinental zone of high productivity primarily coincides with the near-continent region. Since the extent of primary production and the productivity of all the subsequent links in the food chain in the circumcontinental zone average much higher than even in the most productive latitudinal zones of the pelagic region, we arrive at a paradoxical conclusion: the biogenic process typomorphic of pelagic lithogenesis should take place with greater intensity in the nearcontinent region rather than the pelagic. In my view, this paradox can only be solved by assuming that the pelagic biogenic process during the initial stage of sedimentogenesis is universal for all zones of deep ocean, however, it is suppressed by more intensive terrigenous sedimentation in the nearcontinent region. The latter form of sedimentation plays a role in the pellet mechanism for settling terrigenous suspension, but a significant (under conditions of continental slopes and feet, it is the predominant) portion of the terrigenous material is transported in deep and bottom waters lacking biologically active surface layers.

The transformation of sedimentary material in the water column includes, in addition to the described processes biogenically generating suspended particles, a reverse process for their breakdown and dissolution. Actual investigations of sediment flows found that the majority of organic matter and biogenic silica, and a portion of the CsCO_3 produced from nanoplankton in surface waters breaks down and passes into solution in the biologically active surface layers of the water column (at depths of less than 1000 m), returning and circulating their constituent components). This process of dissolution and return to biogenic circulation continues on over the course of settling of biogenic particles through the water column [5, 8, 12].

Pelagic sedimentation includes slow sinking of fine suspension, a situation in which relatively large biogenic particles and fecal pellets produced in the surface waters overtake the fines, chemical precipitation of iron hydrates and manganese, and sorption of microelements upon the latter. The question of the relative settling rates of finely dispersed suspension versus settling of large particles can be answered in favor of the latter in the light of new data obtained with the aid of sedimentation traps. It is precisely this process which controls the distribution of absolute masses of both biogenic and terrigenous components of sediment in the pelagic region. The rapid settling of large particles explains the fact that not only the field of bioproductivity of surface waters, but also biogeographic zones and biocenoses of plankton, patterns of concentration, and the mineral composition of the terrigenous suspension project to the bottom without significant deviations [9, 12].

Within the pelagic regions, three latitudinal belts of high bioproductivity and simultaneously high sedimentation rates (the equatorial belt and two temperate belts) are separated by low-productivity regions with extremely low sedimentation rates under centers of subtropical anticyclone circulation of currents (convergent halystases in the terminology of N. M. Strakov [24, 25], arid zones in the terminology of A. P. Lisitsyn [8, 10]). The concentration of terrigenous suspension and its lateral transport by surface currents, of course, also have an effect upon the distribution of absolute masses of the abiogenic fraction of pelagic sediments, but it is impossible to evaluate the contributions of its variations on the basis of existing data, for high concentrations of suspension in the equatorial zone and at the periphery of the pelagic region [8] coincide with zones of high-intensity biofiltrational settling of this fraction, while minimal concentrations coincide with oceanic deserts, where biogenic processes slack off sharply. The direct correlation between amounts of primary production (as measured by the intensities of all biogenic processes) and accumulation rates of pelagic sediments, including accumulation of their abiogenic fraction, nevertheless indicates a clear dominance of the role of biofiltrational settling outside a direct dependence upon concentration of abiogenic suspension in the surface waters.

Thus, the sedimentation stage in pelagic lithogenesis is distinguished by dominance of a biogenic (bioextractional and biofiltrational) mechanism which is governed by patterns of surface-water bioproductivity conforming to a latitudinal zonation in pelagic regions. The zonation of bioproductivity prevents the development of latitudinal megafacial zones with high and low absolute masses of both biogenic and abiogenic components of pelagic sediments. Each of the zones is characterized by its own specific set of lithological-facies types of sediments and is marked by certain facial attributes [14]. The biogenic mechanism of

sedimentation dominates at relatively low rather than high (relative to near-continent lithogenesis) bioproductivities of waters. The uneven, zonal character of the distribution of absolute masses of various abiogenic components of pelagic sediments is primarily controlled by biogenic factors rather than routes of transport by currents.

Chemogenic (and biochemical) precipitation of dissolved components from bottom waters resulting in pelagic ore formation should be assigned to a subsequent, very important and specific stage in pelagic lithogenesis encompassing the totality of processes at the water—ocean floor interface.

Processes at the surface of the ocean floor play an especially important role in pelagic lithogenesis because of the low sedimentation rates, which promote prolonged exposure of newly precipitated sedimentary material to the action of well aerated bottom waters. The effect of prolonged exposure is intensified by the temporal and spatial heterogeneity of pelagic sedimentation and the presence of numerous areas of exposed ocean floor and zero sedimentation, expressed in sections in the form of stratigraphic breaks.

Processes at the water—sediment interface are an intermediate link between two main stages in lithogenesis: sedimentogenesis and diagenesis; but these processes are fairly specific, so they can be identified as a separate stage. This is a geochemical barrier between two environments with different physical-chemical properties, bottom water and watersaturated highly porous sediments. The ocean-floor surface is also a dynamic barrier separating moving bottom waters with sedimentary material suspended within them from immobilized muddy waters filling the pore space of sediments fixed on the bottom. In other words, the bottom surface serves as a boundary between a very dilute suspension obeying the laws of hydrodynamics (a Newtonian liquid) and a biphasic dispersion system exhibiting the properties of a viscous liquid or solid body governed by rheological laws. Advective transport and turbulent mixing of waters is actively taking place in the near-bottom layer above the barrier, whereas we primarily encounter diffusion (molecular, colloidal) and also filtration of pore solution below the boundary.

The route of sedimentary particles downward through the water column under the influence of gravity ends at the bottom surface, and prolonged interaction (dynamic, physical-chemical, and biochemical) of solid phases of the sediment with bottom water begins. It is precisely the duration of this interaction, a time inversely proportional to the sedimentation rate, which is the main peculiarity of this stage of pelagic lithogenesis as compared to near-continent lithogenesis.

The time-in-residence of a particle on the bottom surface (prior to burial) depends upon the particle's size. Thus, residence equals 10^2 - 10^3 years for particles on the order of 0.1 mm (foraminifers, radiolarians, pellets) with average sedimentation rates ranging from 1 to 10 mm/1000 years; for particles on the order of $1\ \mu\text{m}$ (the thickness of a shell part from a diatom coccolith), it equals 1-10 years respectively. On an average, the duration of direct contact of particles with bottom water (under a standard sedimentation regime) is two-three orders of magnitude longer in the pelagic region than in the near-continent region. The duration of the contact stage differs from the duration of the sedimentation stage by approximately the same magnitude. This is precisely why interactions at the water—ocean floor barrier are of the character of a completion, in contrast to the near-continent region, where rapid burial prevents completion of these interaction.

Interaction at the water—ocean floor barrier includes lithodynamic processes of repeated roiling, erosion, hydrodynamic and gravitational reworking of sedimentary material, biogenic carbonates and silica), and authigenic minerogenesis by precipitation from water washing the bottom (hydrogenic precipitation) and as a result of downward diffusion from muddy solutions. In exposed areas, halmyrolysis (submarine weathering) of igneous and metamorphic rocks of the basement and hyaloclastics of submarine eruptions are added to these essentially pelagic processes; at active fracture zones, there is also tectonic crushing of rock, with subsequent transport of the edaphogenic material formed and entry of this material into the composition of pelagic sediments [13, 14]. Although products of halmyrolysis, edaphogenic materials are not, strictly speaking, classified as pelagic components, they should be considered natural to certain pelagic facies all the same [14], and thus natural to pelagic lithogenesis in general. Submarine hydrothermal springs discharge water into the bottom layer, thereby also making a contribution to pelagic lithogenesis, but according to patterns of volcanogenic—sedimentary lithogenesis.

The fact that pelagic sedimentation is not a random, one-time sinking of a "rain" of particles to the bottom, followed by burial at the site of incidence has been known for a long time, but the scale and nature of bottom redistribution of sedimentary masses in the pelagic region under the influence of hydrodynamics and gravitational mixing has yet to be understood.

In pelagic sedimentogenesis, many lithodynamic types of sediments analogous to those occurring in near-continent regions are formed [14], including: sediments of bottom currents (karrenites), clouds of turbid water (nepheloidites), turbidites, and various gravitational formation (gravities); however, their relative role and form of manifestation differ essentially, not to mention the different composition of the original material. The first, and probably the foremost difference consists in the narrow localization of lateral sedimentological flows and small masses of sedimentary material incorporated in them. Turbid flows running off the continental slopes travel hundreds and thousands of kilometers, and sediments of a single mudstone flow can

have a thickness of hundreds of meters; in the pelagic zone, all these processes develop within local rises and depressions. The sedimentary material transported by them is exclusively local (pelagic biogenic or edaphogenic); the thickness of sediments in a single flow is usually in the tens of centimeters.

Pelagic gravities and turbidites are distributed on the flanks of midocean ridges, where the sharp dissection of relief and seismotectonic activity favor their development. Here, one finds small sedimentary basins, "dams" filled with carbonate bioturbidites up to 500 m thick or more accumulating at an average rate of tens of millimeters per 1000 years [2, 14, 17]. Volcanogenic-edaphogenic taluses consisting of fragments of pillow basalt have been described in rift valleys, and edaphogenic gravities and turbidites [13, 14, 17] have been described at active transform faults. Processes of gravitational reworking are much less well developed within intraplate tectonically little-active regions of the ocean floor, where they coincide with regions of sharply dissected relief exhibiting sharp slopes and large drop depths: aseismic ridges, fracture zones, and seamounts. The gravities here are predominantly edaphogenic clumpy and crushed taluses and also accumulations of ferromanganese nodules which rolled along the slope.

Little is known concerning manifestations of slow gravitational creep of the upper semiliquid layer of sediments in pelagic regions of the ocean, although many investigators assign a prominent place to its action, including action connected with the formation of dense accumulations of ferromanganese nodules on gentle slopes of abyssal banks. The mechanism behind the formation of the layers and lenses of *Etmodiscus* muds is yet to be discovered. Their coincidence with local depressions and also the composition and certain textural features suggest reworking.

The chief and most universal mechanism involved in the reworking of pelagic sediments is the hydrodynamics of bottom waters. Roiling of sediments, washing them of fine fractions (formation of foraminiferous sands by washing coccoliths at the tops of seamounts), erosion of the bottom, transport in a suspended state and settling when the current velocity drops take place under the influence of hydrodynamic agents. In contrast to continental feet, where bottom "contour" currents constantly acting and following along isopaths transport and deposit huge masses of terrigenous material; the activity of currents in the pelagic region, like other reworking processes, is not expressed in the form of extended zones; it is localized within individual morphostructures or their parts.

So-called *benthic storms*, abrupt impulses of high hydrodynamic activity of bottom waters resulting in roiling of the surface layer of sediments and the formation of nepheloid clouds above the bottom, probably have special importance in the redistribution of pelagic sediments. Little is known concerning abyssal benthic storms in the pelagic region, but their possibility is acknowledged by the majority of investigators (for example, [29]). During a month-long series of measurements of bottom currents on a rolling abyssal plain in the northern near-equatorial zone of the Pacific Ocean, one short impulse of high (up to 15 cm/sec) current velocity was recorded [6].

The most important dynamic agent of pelagic lithogenesis at the stage of water-ocean floor contact is *bioturbation* the biomechanical reworking of sediments (and, in some cases, lithified rock also) by benthic organisms. There is too extensive a literature on the subject of bioturbation of pelagic sediments to be reviewed in this article. I will only emphasize certain issues which are important to the problem under consideration here. Firstly, the intensity of bioturbation depends upon the sedimentation rate; when accumulation is slow, the activities of separate, even rare individuals add up" over the long period of time the surface layer resides in their zone of habitation.

Secondly, bioturbation manifests not only in the form of clear tracks (burrows) of burrowing organisms prepared to some degree by early diagenetic redistribution of iron and manganese and by creation of completely mixed unbedded homogenized textures characteristic of the surficial semiliquid layer of pelagic muds.

The nature of the homogeneous semiliquid surface layer of pelagic sediments, well known to all who are engaged in first-hand description of modern sediments, was investigated by V. Berger [28]. Based upon the example of carbonate sediments of the Ontong-Java rise (the western preequatorial portion of the Pacific Ocean), he demonstrated that there is no vertical gradient of ^{14}C in the upper (~ 6 cm) layer homogenized as a result of bioturbation, i.e., "age-averaging" of sediments exists within a certain time interval, in the actual case under consideration, within the Holocene. It is evident that the time interval encompassing biogenic homogenization is inversely related to the sedimentation rate. Therefore, it is hardly possible to extrapolate evidence of a Holocene age of an homogeneous layer obtained for carbonate sediments with high sedimentation rates to all pelagic sediments [18].

The role of bioturbation is not limited only to mechanical mixing of sediments, resulting in their textural homogenization. As a result of the activity of burrowing organisms, especially mud-eaters, organic matter not completely broken down is "covered" in the sediment with a surficial film; this promotes early diagenetic activity of the homogeneous layer, including partial

reduction of manganese oxyhydrates and their diffusional migration to the surface. On the other hand, mud-eaters carry already successfully transformed muddy solutions to the surface, maintain the semiliquid constitution of the muds, and perform roiling of the sediments, enriching the bottom waters with a second suspension and promoting reworking processes.

Gradients of contents of C_{org} , O_2 , and a number of other components were discovered in a homogeneous surface layer during detailed testing [32]. This indicates that the processes taking place in the layer are of a dynamic character. At the same time, V. N. Sval'nov [18], studying variations in the thickness of a semiliquid layer in a polygon in the central basin of the Indian Ocean studied in detail, was led to conclude that there is mechanical mobility of this layer and that it is subjected to rewashing, even under the influence of weak bottom currents. The inconstancy and ephemeral nature of the semiliquid homogeneous layer, also supported by our observations [22], together with the data already presented on bioturbation, leads the author to the conclusion that the homogeneous layer of semiliquid mud belongs to the water—ocean floor contact zone rather than to a separate stage of early diagenesis. In this same approach, as already mentioned [16], an essentially unified pelagic ore process, the formation of ferromanganese nodules, can be placed entirely within the stage of water—ocean floor contact specific to pelagic lithogenesis. That being the case, the classification of early diagenesis, in the understanding of N. M. Strakov, as a separate stage (or substage) loses its overall meaning. N. M. Strakov himself [23, 25] wrote about the weak manifestation of early diagenesis (and diagenesis in general), believing this to be one of the features of the oceanic type of lithogenesis he was identifying.

The water—ocean floor contact is a physical-chemical barrier at which complex and multidirectional reactions of authigenic minerogenesis, solution, and diffusional and filtrational exchange between muddy and bottom waters takes place. The geological effects of these processes under otherwise equivalent conditions are more clearly manifest the longer a given contact surface remains unburied under new portions of sediment.

Below a certain depth referred to as the critical depth (or carbonate compensation depth, CCD), reactions of complete dissolution of biogenic particles of $CaCO_3$ or their aggregates (pellets filled with coccoliths) take place over the entire ocean-floor surface. In the process, microelements confined in the carbonate are liberated and passed into bottom or near-surface muddy water in the form of free ions or colloids. It is believed that settling while within the composition of carbonate biogenic particles is the main mechanism of transport of iron and a number of other metals in pelagic sediments and also in ferromanganese nodules and crusts [30, 31].

The pattern of critical depth of carbonate accumulation discovered on the *Challenger* expedition [33], universal for deep-water sedimentation, is realized in pure form only in pelagic lithogenesis; whereas its action is often masked by intensive accumulation of terrigenous material in the near-continent regions. Since biogenic carbonate production by planktonic microorganisms takes place (although at different intensities) over the entire pelagic region of the ocean, while the sedimentation flow of $CaCO_3$ apparently everywhere exceeds the absolute masses of flows of other biogenic components, carbonate-free pelagic sediments can form only below the CCD (for complete solution of $CaCO_3$), being the insoluble carbonate-free remains of the total sedimentation flow. At lesser depths, pelagic sediments are ubiquitously carbonate (with the exception of areas of erosion or nondeposition, of course). It follows that all carbonate-free pelagic sediments are deep-water in character, relative to neighboring carbonates.

The facial transition from carbonate-free pelagic sediments to carbonate sediments (at lesser depths) has an uneven, stepwise character; it passes through a series of successive critical dissolution levels, lysoclines, and critical depths for various biomorphic modifications of $CaCO_3$ [3, 5, 9, 14]. In combination, these critical levels form a depth-determined facial solution series universal for pelagic lithogenesis [14]. The stepwise character of the variation in carbonate solution rates can be explained, on the one hand, by different solubilities of various sediment-forming biogenic remains, aragonitic shells of pteropods and heteropods, calcitic shells of planktonic foraminifers of different species (it is precisely the classification of the latter into resistant and unresistant to solution which serves as a basis for delineating the foraminifer lysocline [31]), and coccoliths. On the other hand, the critical levels are controlled to some degree by the vertical structure of water masses, especially the position of the upper boundary of Antarctic bottom waters "aggressive" with regard to carbonates. Since biogenic carbonates dissolve predominantly at the surface of the ocean floor, dissolution is affected by water masses coming into contact with the ocean floor at various depths, rather than the stratified water column itself through which the carbonate remains of planktonic organisms settle toward the bottom.

Under conditions of pelagic lithogenesis, the major portion of planktonogenic silica (diatoms, radiolarians, and silicoflagellates) are dissolved at the surface of the ocean floor; but, in contrast to $CaCO_3$, the dissolution effect is controlled not by depth, but by rate of arrival and rate of burial. Solution continues, although more slowly, after burial, during the stage of diagenesis, and, possibly, during catagenesis.

Less than 90% of the organic matter reaching the bottom, and almost all of its labile energetic components, break down in the most surficial layer of sediments under the direct action of well aerated bottom waters. Only an insignificant portion of organic matter produced in the zone of photosynthesis, represented entirely by the stable component kerogen, is buried in pelagic sediments; this reduces the diagenetic activity of organics to zero and provides an oxidizing environment of diagenesis characteristic of pelagic lithogenesis.

At the geochemical water—ocean floor barrier, where biogenic particles settled through the water column are kept for a long time prior to their burial, these particles are thus subjected to thorough chemical and biochemical reworking. As a result of breakdown and solution, certain chemical components are liberated from them; depending upon the chemical properties of the bottom waters under stable physical-chemical conditions, they either pass into solution and return to circulation or fall into the sediment in the form of authigenic mineral species.

An ore process specific to pelagic lithogenesis, formation of authigenic oxyhydrates of iron and manganese in the form of nodules, crusts, microconcretions, and finely dispersed forms disseminated in pelagic sediments take place at the geochemical water—ocean floor barrier. In addition to Fe and Mn, more than ten chemical elements, the most important being Cu, Mi, Co, and also Pb, Zn, Mo, and Pt, are concentrated in the ore matter of the concretionary formations [1, 19, 21]. Such a set of ore components, their quantitative ratios, and concentration coefficients (relative to clarkes of pelagic sediments) are characteristic only of pelagic lithogenesis and are its distinctive attributes.

On the basis of morphological, structural-textural, mineralogical, and geochemical attributes, we can clearly distinguish two main types of oxyhydrate ore material, referred to sedimentational (or hydrogenic, i.e., precipitated directly from bottom water) and diagenetic, for which precipitation from muddy waters in the most surficial layer of enclosing sediment is assumed. Thus, processes in the semiliquid geochemically rather active surficial layer are assigned here to the stage of water—ocean floor contact; the term *diagenetic* is used conditionally. But the conception of postsedimentational remobilization and diffusional migration of at least manganese through muddy water to growing "diagenetic" nodules (from below and, apparently from the side primarily) is used as an explanation for the sharply increased values of Mn/Fe [19-22].

Process occurring at exposed surfaces of ancient sediments or bed rock of the ocean floor which result in the formation of a "submarine eluvium" of its own kind should also be assigned to the stage of water—ocean floor contact. These are halmyrolytic alterations of volcanic rocks (especially hyaloclastites), metasomatic phosphorites of seamounts, the ferromanganese crusts already mentioned, and various lithification crusts (hardgrounds) and their incipient manifestations in the form of zones of "ferrugination" and compaction, which mark surfaces corresponding to stratigraphic breaks in the sections. Pelagic hardgrounds were described in detail in the work of H. Jenkins [4] his work relieves us of the necessity of addressing this issue. The fact that hardgrounds, previously considered to be exclusively shallow-water formations, arise in pelagic lithogenesis at any depth down to abyssal was recently discovered by marine geologists; but this disturbing fact, like the existence of numerous breaks in sedimentation in the open ocean, was not given much importance until recently.

The submarine—elluvial process develops in one or another fashion under a single condition: prolonged exposure of an outcropping surface to the action of bottom waters. The character of this process and its material (geochemical, mineralogical) expression depends, on the one hand, upon the composition of the substrate and, on the other, upon the properties of water masses washing the bottom, including the saturation of the water masses with oxygen, carbon dioxide, phosphorous, and various microcomponents. Thus, it is assumed that location of the exposed surface of the peak of a seamount directly beneath the lower boundary of an intervening layer of minimum oxygen in the water column facilitates the formation of ferromanganese crusts and their enrichment with manganese and cobalt [26, 31]. Metasomatic phosphorites of seamounts probably formed within the layer of an oxygen minimum during periods of its especially intensive development [4]. For the formation of carbonate (usually magnesian calcitic) cementation crusts in pelagic calcareous sediments, the waters washing the bottom should be oversaturated with carbonate.

Diagenesis in the pelagic sedimentary process is distinguished, first of all, by the fact that it takes place entirely under oxidizing conditions without a major source of energy, breakdown of organic matter. Therefore, redox reactions appear only very weakly and especially locally (for example, around the burrows of mud-eaters) and, apparently, the diagenetic migration of elements of variable valence caused by these reactions, a migration characteristic of marine humid lithogenesis according to the reports of N. M. Strakov, I. I. Volkov, and a number of foreign authors, is practically absent. N. M. Strakov, having in mind precisely pelagic lithogenesis in our understanding of the word, writes, "the stage of early diagenesis is indistinguishable here from the stage of late diagenesis, and even early catagenesis" [25, p. 195]. As noted above, early diagenesis in our conception is rather merged with the stage of water—ocean floor contact. Its occurrence is barely noticeable even in the top several decimeters of the section of Quarternary pelagic sediments, where changes of several physical-chemical parameters of muddy waters, the

content of oxygen and nitrates, pH and alkalinity, and also the content of organic carbon in the sediment can be expressed in the form of exponential curves [32]. The lithological consequences of these changes are apparently insignificant.

Under such conditions, processes of diagenesis are predominantly isocheimic, consisting in gradual decreases in moisture and porosity, compaction of sediments, recrystallization, syneresis of colloids, and the emergence of several neomineralizations. Examination of core from hundreds of deep-water boreholes showed that these processes proceed differently in three main groups of pelagic sediments: carbonate, clayey, and siliceous. In carbonate sediments, recrystallization of primary biogenic calcite results in the formation of a porous framework and a conversion of soft calcareous mud to semilithified porous fine-grained limestone. Pelagic clays are compacted and acquire rigidity; clay minerals are partially recrystallized and subjected to syneresis, and ferruginous-aluminiferous gels, which were probably present in the original sediment, are crystallized into authigenic ferrismectites and zeolites. In siliceous sediments, dissolution of siliceous organic remains and partially redistributed silica continues; biogenic amorphous silica (opal A) crystallizes with the formation of disordered cristobalite-tridymite (opal CT), which replaces biomorphic particles, and precipitates in the pore space in the form of lamellar-spherulitic structures: lepispheres.

Diagenetic silica in the form of conformably bedded lenses and stringers occur in all types of pelagic sediments. According to deep-water drilling data, they usually occur in layers no younger than Eocene, although occurrences of silics in the Neogene are also known. In Paleogene sediments, the silica is usually cristobalitic-tridymitic (opal CT), porcellanites, and, in the Cretaceous and Lower-Jurassic, quartz-chalcedony rocks.

Diagenetic clinoptilolite, which is apparently paragenetically associated with opal CT, frequently occurs in various types of pelagic sediments; when the latter is converted to chalcedony and quartz, the clinoptilolite usually disappears.

The conversion of nano-fines to sturdy microcrystalline limestones with complete disappearance of biomorphic structure the conversion of tridymitic-cristobalitic porcellanites to quartzic silicas, the conversion of dense clays to unsoaked argillites, and the conversion of goethite-hydrogoethite to hematite, which is found in the Early Cretaceous in the example of sections of the northwestern portion of the Atlantic [15], apparently should be considered a transition to diagenesis to catagenesis.

* * *

The consideration presented above allow us to conclude that the pelagic sedimentary process taking place in open areas of large, tectonically "mature" ocean basins, due to their own specific characteristics at all stages (from preparation of the sedimentary material to conversion into sedimentary rock) can be classified as a special type of lithogenesis. The proposed conception of pelagic lithogenesis cannot be reduced to the oceanic type of lithogenesis delineated by N. M. Strakov. The Strskovishn oceanic sedimentary type of lithogenesis is only a part of this conception.

The following are key stages in pelagic lithogenesis: 1) the stage of pelagic sedimentogenesis, where, in contrast to all other types of lithogeneses, a biogenic mechanism governing the biological productivity of the surface waters of the ocean predominates; 2) processes at the water-ocean floor contact, where prolonged mechanical and chemical interaction of sedimentary material fallen to the bottom or outcropping rock with bottom water takes place. During both of these stages, a complex transformation of the original material is accomplished, with the majority of it returned to circulation and only a minor portion passing on the next stage. The outcome of cycles is of a strictly selective character, finally resulting in a thorough differentiation of the sedimentary material and determining the geochemical specifics of pelagic lithogenesis, including sedimentary ore genesis.

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ASPECTS OF LATE QUATERNARY SEDIMENT FORMATION IN THE BLACK SEA PART 2: FORMATION OF REACTIVE IRON COMPLEXES

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Characteristics of the distribution of certain reactive forms of iron within marine deposits, described in Part 1, have been analyzed. These include pyrite, hydrotroilite, nonsulfide forms of Fe(II) and Fe(III), and also organic carbon. An interpretation of the data is proposed.

The results of an investigation into the distribution of reactive forms of iron in Late Quaternary deposits of the Black Sea, presented in Part 1, together with data from earlier studies, are discussed in terms of the extent to which they are consistent with existing theories of the geochemical evolution of the sea during the Holocene. These theories, formulated by Strakhov on the basis of a detailed analysis of all the lithological and geochemical material available to him at the time, were largely concerned with the origin of the Ancient Black Sea muds, the sulfide (hydrotroilite) mineralization of Neoeuxinic deposits, and the contamination of the water column by hydrogen sulfide. These phenomena were regarded as being related to a single process [14, 15].

The theory proposed by Strakhov also explains a wide range of physical, chemical and biological phenomena, and is based mainly on the fact that there is a chronological correlation between the timing of the breakthrough of Mediterranean salt water through the Bosphorus and the onset of intense bioproductivity, which led to the development of sapropelic muds in the Ancient Black Sea deposits. The abrupt increase in bioproductivity is thought to have resulted from the movement of biogenic components from within the denser water in the deeper part of the basin into the photic zone. The increased sedimentation of organic matter and the simultaneous enrichment of the bottom waters with sulfate from the Mediterranean, according to this theory, led to the development within the Black Sea sediment column of a horizon with an exceptionally high level of sulfate reduction. This led on the one hand to H_2S migrating into the water column, and on the other hand to its diffusive flow into the deeper deposits. The interaction of H_2S with the ferruginous Neoeuxinic sediments typically resulted in hydrotroilite mineralization. In addition to diffusive flow of truly soluble components (H_2S , HS^- , SO_4^{2-} , Fe^{2+} , Cl^- , Na^+ , Mg^{2+}), this theory proposes the diffusive movement within the sediment of newly formed colloidal-dispersed phases: Fe(II) monosulfides; stabilized H_2S ; and also elemental sulfur. It is proposed that the migration of H_2S into the overlying waters, and its convective redistribution within the lower, more dense zone of the stratified water column, resulted in the gradual raising of the density boundary and the formation of the modern hydrogen sulfide zone within the sea.

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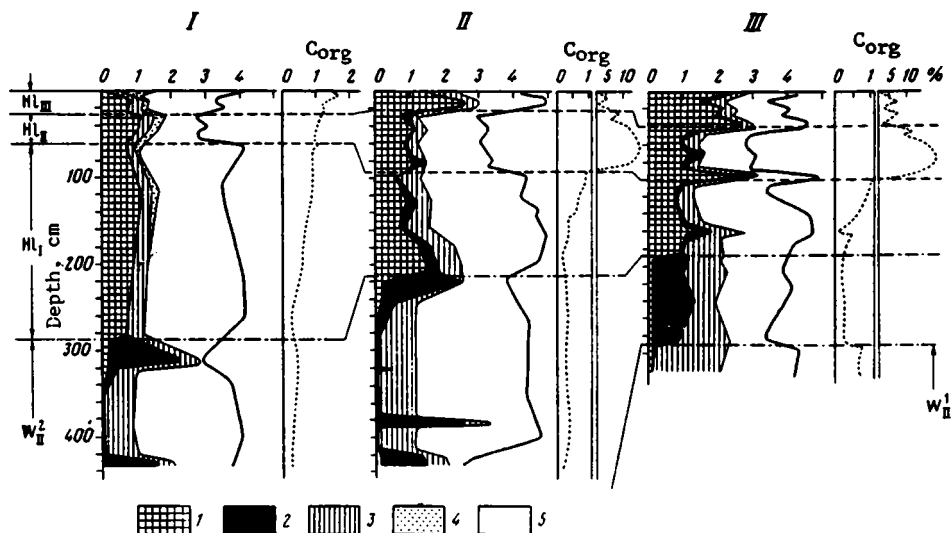


Fig. 1. Main characteristics of the distribution of forms of iron within the Late Quaternary sediments of the Black Sea (generalized). I) Upper slope and lower shelf (oxygenated zone); II) Lower slope and base of slope; III) pelagic zone of western Khalistaz. Stratigraphy: W_{II}^1) early Neoeuxinic; W_{II}^2) middle Neoeuxinic (hydrotroilitic); H_{I1}) late Neoeuxinic; H_{II1}) Ancient Black Sea; H_{III1}) modern. Forms of iron: 1) Fe pyrite; 2) acid-soluble Fe sulfides (hydrotroilitic); 3) nonsulfide reactive Fe(II); 4) reactive Fe(III); 5) detrital forms.

This accepted model does not raise any doubts in relation to the main proposal; that biogenic components were displaced into the photic zone and led to the formation of the Ancient Black Sea sapropels, in so far as the circulation of these components forms an essential condition for any occurrence of the processes of marine bioproductivity. In the Black Sea the organic reserves required for bioproductivity were present, together with all the other requirements for it to occur. However, with a detailed examination, a number of serious objections can be raised against the current theory. The first of these concerns the question of whether the Ancient Black Sea sediments could really have formed such a major and long-lived source of bacterial H_2S , and also the possibility of the diffusive mechanism having been responsible for the formation of Neoeuxinic sulfide mineralization, and for the contamination of the water column by hydrogen sulfide.

A discussion of these problems can best be tackled in the order in which data were presented in Part 1 of this paper; i.e. through the sedimentary section from the base upwards, using the provisional subdivision of the Neoeuxinic sediments into early, middle and late Neoeuxinic horizons (W_{II}^1 , W_{II}^2 , and H_{I1}) [18]. For convenience the main characteristics of the distribution of the components studied have been represented graphically in a generalized form (Fig. 1).

NEOEUXINIC DEPOSITS

It is evident that the total accumulation of reduced sulfur within the sulfur-enriched Neoeuxinic deposits on the upper slope and shelf is on average not lower, and in some cases is substantially higher, than in the corresponding deep-water sediments (Fig. 1). This is clearly demonstrated by the almost average content of pyrite within the sediments of the late Neoeuxinic horizon (H_{I1}), which is appreciably thicker in the upper parts of the basin. Allowing for the same average content of sulfides in the dispersed phase of the sediments of the hydrotroilitic horizon, and its thickness, the same assertion may be made for the sulfidized succession in the Neoeuxinic sediments as a whole (Fig. 1). Furthermore, the Neoeuxinic sediments on the slope (including its upper part) were enriched significantly in comparison with the pelagic zone by iron sulfides in the form of semiconsolidated clots, concretions, crusts and pyritized organic detritus [5, 7].

Assuming that diffusion of the type proposed occurs, then the nature of the distribution of the Neoeuxinic sulfides suggests that the sulfate-reducing process in the Ancient Black Sea sediments is just as intense in the upper part of the slope and shelf as in the pelagic zone. This is highly improbable. It is inconsistent with the presence within the late Neoeuxinic and Ancient Black Sea sediments of a number of regions in the upper part of the basin with a significant amount (10-25% of total reactive Fe) of reactive nonsulfide trivalent iron (see Part 1 of this paper), which would not be preserved in the presence of an

intense diffusive flow of H_2S . At the same time, the existence of such a flow would inevitably suggest a much more intense flow of H_2S was taking place into the water immediately above the sea bed, i.e. a protracted steady contamination of the water column on the shelf, which is completely ruled out by the active hydrodynamics of the marine coast line. Of course, the possibility of the periodic onset of anaerobic conditions, or reduced-oxygen conditions within the waters on the shelf cannot be dismissed during periods of intense Ancient Black Sea bioproductivity. These circumstances cast some doubt on the proposed diffusive mechanism for the formation of the Neoeuxinic sulfides in sediments within the upper part of the basin, and therefore also for those in the deeper areas.

Another problem with the diffusive theory is the typically stratified nature of the hydrotroilite on the slope and shelf. Within the middle Neoeuxinic horizon it is concentrated within individual beds, separated by a significant thickness (up to 60 cm) of sometimes virtually sulfide-free sediments (Fig. 1) [13, 18]. It is very doubtful that these sediments, which have quite high contents of reactive Fe(II) (1-1.5%), remained unaffected by a diffusive flow of H_2S . This may have been due to the formation of forms of Fe(II) which were inaccessible to H_2S , but the existence of beds exclusively of these forms, with lithologically identical sulfidized interbeds, is difficult to explain. Attempts to explain the observed stratification by the diffusive flows encountering soluble Fe^{2+} , or the redistribution of colloidal-dispersed FeS in pore waters, are also unconvincing [5, 13]. The first of these is invalid, since the solubility of iron monosulfides is similar to or higher than that of silicate or aluminosilicate compounds of Fe^{2+} (solubility of FeS = $10^{-18.4}$, solubility of FeSiO_3 = $10^{-18.9}$) [6]. The migration of colloidal FeS is inconsistent with the very restricted diffusive movements of the colloidal-dispersed particles which comprise their basic physical structure. Furthermore, even the minimal salinities measured within the pore fluids of the Neoeuxinic sediments (4 g/liter of Cl^-) do not allow the stabilization even of significantly more stable colloids such as $\text{As}_2\text{S}_3 \cdot n\text{H}_2\text{S}$, the threshold of coagulation of which is known to be 51 mmole/liter, or 1.81 g/liter of Cl^- [18].

Therefore, the existence of isolated hydrotroilitized interbeds is also incompatible with a diffusive model.

Similar problems arise in explaining the cause of the marked predominance of pyrite in the sulfide mineralization of the sediments in the late Neoeuxinic horizon ($\text{H}\ell_1$), which separates the hydrotroilite deposits from the Ancient Black Sea deposits (Fig. 1). The pyrite is thought to have formed as the result of a secondary process of pyritization of original FeS, due to sulfur derived from the Ancient Black Sea deposits and the bottom waters in soluble form ($\text{S}_2\text{O}_3^{2-}$) [4]. However, stoichiometrically: $\text{FeS} + \text{S}_2\text{O}_3^{2-} = \text{FeS}_2 + \text{S}_2\text{O}_3^{2-}$, so this mechanism requires the diffusive supply to the Neoeuxinic sediments of thiosulfate in a quantity equivalent to monosulfide; i.e. it consumes H_2S . This may not be provided by either sulfate reduction itself, or by oxidation of H_2S , since ions of $\text{S}_2\text{O}_3^{2-}$ are present in the process only in the form of an intermediate product, in highly subordinate quantities in comparison with H_2S [19, 20]. Furthermore, according to the proposed theory, the inevitable contamination of the bottom waters by hydrogen sulfide must separate the sedimentary surface from the zone of H_2S oxidation, which implies that $\text{S}_2\text{O}_3^{2-}$ cannot participate in the diffusive process. Moreover, the inevitable consequence of pyritization must be a marked accumulation within the pore waters of a horizon of SO_3^{2-} ions (together with polythionates), which in these sediments are virtually absent [4]. Therefore it is suggested that diffusion of $\text{S}_2\text{O}_3^{2-}$ and the pyritization of monosulfides, if they take place within the sediments, occur only on a small scale and they are not the main cause of sulfide mineralization.

All these facts are sufficient to cast doubt on the diffusive model for the formation of the Neoeuxinic sulfides, and the role of the Ancient Black Sea muds as the source of hydrogen sulfide. However in recent times additional evidence has been obtained which emphasizes the limited role of diffusion.

An investigation of the chemical and mineralogical composition of the sulfide within the hydrotroilitized sediments has shown that the hydrotroilite forms a complex multicomponent system of genetically interrelated iron sulfide compounds, dominated by sulfides of Fe(III) (greigite, hydroxysulfides), and the monosulfide FeS is present only in subordinate quantities or is absent [9]. It has been shown experimentally that greigite (in addition to FeS) is a decomposition product of the Fe(III) hydroxysulfide compounds, since the relationship between these forms in a sample is characterized by the extensive decomposition of the initial phase and shows the relative age of the sulfide formation. This leads to the conclusion that the sulfide phases in the lower beds of the hydrotroilite horizon are highly enriched in greigite which formed earlier than the phases in the upper beds, where Fe(III) hydroxysulfides dominate. The nature of this distribution obviously contradicts the postulated direction of diffusive flow of H_2S from above downwards, and demonstrates the successive formation of sulfides as the sedimentary succession was deposited.

One more important line of evidence against the existence of diffusive flow of H_2S is given by the results of a study of the isotopic composition of reduced sulfur in the sediments. These analyses were presented in Part 1 of this paper [16]. They show the successive downward increase in the density of isotopes of sulfur from the Ancient Black Sea sediments ($\delta^{34}\text{S} = (-30)$ to $(-40)\text{‰}$) to the Late Neoeuxinic deposits ($\delta^{34}\text{S} = (-10)$ to $(+10)\text{‰}$). This trend continues to the hydrotroilite bed, where $\delta^{34}\text{S}$

reaches +17.8, and even as high as +25.6‰. In our opinion these data have not been convincingly interpreted, but there is no doubt that this distribution of sulfur isotopes could not be the result of a diffusive process, since such a process would produce the very opposite effect.

Some explanation is required to account for the numerous types of data which contradict the diffusive model for the formation of the Neoeuxinic sulfides. This needs a more thorough consideration of all the available material and further field and experimental work. However, it is evident that an alternative approach to this problem is to consider the possibility that sulfate reduction and sulfide formation occurred within the Neoeuxinic basin itself.

The environment of the early Neoeuxinic sea-lake (with a salinity of 8-12‰; $\text{SO}_4^{2-} \sim 0.100$ g/liter; C_{org} 0.72%) would not have prevented the development of microbiological sulfate reduction, although conditions for it do not appear to have been particularly favorable [20]. The most important factor to restrict this process is the presence or absence of particular groups of organic compounds which may be assimilated by microbiological organisms. These are mainly represented by oxidized forms of organic matter [1, 3, 7]. It is therefore logical to suppose that a sharp increase in the intensity of sulfate reduction and sulfide formation in the Neoeuxinic basin may have been related to the onset of well-aerated glacial meltwaters entering the basin. This must have led to the partial oxidation of reduced forms of organic matter, in addition to Fe(II), which had previously accumulated in the basin over an extended period of time. A characteristic feature of such a process is apparently the inadequacy of the oxygen supply compared to the reserves of reduced elements. Therefore, oxygen would have been consumed over a long period of time, providing the possibility for acid-soluble and slightly oxidized Fe(III) sulfides to be formed and preserved, such as hydroxysulfides and greigite [9].

In relation to this it is interesting to note the sediments from Station 840 (see Part 1) in which a reduction in the contents of C_{org} , CaCO_3 , and Fe ions accompanied the occurrence of hydrotroilite. The variations in CaCO_3 and Fe have a clear sedimentary character. It is reasonable to suppose that these variations, together with changes in the lithology and pollen assemblages, reflect a change in the depositional environment, and that this was also directly related to the formation of acid-soluble sulfides. The decrease in C_{org} may be due to the increased consumption of organic matter with the development of the sulfate-reducing process.

A number of hydrotroilite interbeds which are anomalously low in iron, or in which it is completely absent (Fig. 1), favor the proposal that sulfide formation took place within the Neoeuxinic basin itself. These relationships are inexplicable in terms of existing models, since they would not be produced either by weak diagenetic processes in the Neoeuxinic basin, nor by diffusional sulfidization. It may therefore be proposed that the coincidence of maximum sulfide mineralization with the minimum content of detrital forms of iron results from an abrupt increase in oxidation-reduction transformations of detrital forms of iron in the water column, and quite rapid (geologically instantaneous) deposition of a range of reactive forms, including sulfides. This presupposes the possibility of sulfate-reduction and sulfide-formation processes operating on suspended material.

Such a presupposition is perfectly consistent with the observed stratification of hydrotroilite into individual beds. These may reflect on the one hand a periodicity in the input of glacial waters related to periodic warming of the climate during this period. On the other hand they may result from the formation of local mixing zones within the water mass. Either of these could result in distinct stratification in the sediments on the shelf and slope. Such stratification may be absent in pelagic zones where a slower and more uniform deposition of fine sediments, including sulfides, took place.

Evidence of sulfate reduction and sulfide formation may also be seen in the sediments of the late Euxinic (H1) horizon in deep-water parts of the sea. For example, in individual beds there is a convincing relationship between sharp changes in residual C_{org} and rhythmic changes in iron sulfides (Fig. 1). The significance of these relationships is still not fully clear, but the observed effect is most probably due to changes in the diagenetic environment.

It should be emphasized that the interpretation proposed here is conjectural, and has yet to be confirmed.

THE ANCIENT BLACK SEA DEPOSITS

The data presented above, suggesting that the diffusion of H_2S is not responsible for the formation of the Neoeuxinic sulfides, leads to the consideration of the distribution of the composition of the Ancient Black Sea sediments, and particularly from the viewpoint of the proposed role of these sediments as an important and geologically long-lived source of hydrogen sulfide.

The basis for this idea was the presence of sapropels within the deep-water marine sediments, i.e. of abundant organic matter which has not been consumed during diagenesis, and which is not at present a source of hydrogen sulfide. This is well illustrated by the data on the distribution of H_2S within pore waters of the sediment column [7]. This already demonstrates the limitation on sulfate-reduction processes in the presence of abundant organic matter. Therefore the high content of residual C_{org} does not indicate the degree of sulfate reduction and H_2S productivity in deep-water deposits during Ancient Black Sea times. Neither is it the basis for supposing that the Ancient Black Sea deposits in the upper part of the basin were a major source of H_2S . These were relatively impoverished in C_{org} (1-2%), and sapropels here were either absent or very poorly developed and, as shown above, in a number of places nonsulfide forms of Fe(III) occurred (Fig. 1).

It is useful here to consider the distribution of reactive forms of iron in the deep-water sapropelic muds. Analytical data show that, despite the generally high level of sulfidization of reactive forms of iron, a significant proportion (up to 40% of total reactive Fe) is in the form of nonsulfide Fe(II). At the same time, total reactive iron within individual beds of some sections is virtually all in the form of iron sulfides (Fig. 1).

The presence of nonsulfide Fe(II) within the Ancient Black Sea deposits has long been known, and has been interpreted as a result of the formation of authigenic iron silicate phases which did not react with H_2S [6, 12]. Such an explanation is justified for sediments in which processes of sulfate reduction were only weakly developed, with only limited amounts of sulfur dioxide generation. However, it cannot be applied to the situation proposed here, in which sulfate reduction was extremely intense, and occurring at the same time as the initial deposition of the Ancient Black Sea sapropels. In this situation, in conditions of abundant sulfur dioxide, divalent iron must be exposed to H_2S immediately it forms, i.e. at the very moment the Fe(III) minerals are broken down. Indeed, the breakdown is primarily the result of H_2S . It is also well-known that nonsulfide forms of Fe(II) which originate in these sediments are represented by aluminosilicate phases of variable composition which pass during an extended process of aging into crystalline forms such as leptochochlorites and chlorites [2]. In comparison with sulfide formation, which is virtually complete once the Fe—S bonds have formed, this process is kinetically retarded since it depends on an ordering of the molecular structure. Therefore the stability of newly formed aluminosilicates containing Fe(II) in the presence of H_2S depends both on their ordering, and on the concentration of H_2S and pH values. The concentration of free H_2S within the pore waters of the Ancient Black Sea sediments, according to the present model, must have been quite high and, in particular, sufficient for the proposed diffusive penetration to the depth of the base of the hydrotroilite horizon, i.e. comparable with the concentration of macrocomponents (SO_4^{2-} , Cl^- , Na^+). This requires a marked decrease in pH values (5.5-5), in the presence of which the sulfidization not only of reactive iron is inevitable, but also (according to experimental data) of a number of crystalline, detrital forms [21, 22].

Therefore, the presence of a substantial quantity of nonsulfide Fe(II) within the Ancient Black Sea sediments demonstrates that sulfate reduction did not occur to any great extent, and that it was insufficient for the sulfidization of all of the reactive forms of Fe. This is emphasized by the clear signs of a positive correlation between iron sulfides, total reactive Fe, and total iron in sections through the horizon (Fig. 1). At the same time, convincing confirmation of this conclusion may also be seen in the existence of individual fully sulfidized beds, demonstrating the availability of reactive iron in these sediments to the activity of H_2S , and apparently reflecting the geologically brief intervals in which its concentration was increased.

It is evident that in this case, nonsulfide forms of Fe(II) did indeed exist within the aluminosilicate phases, formed in conditions of moderate development of sulfate reduction and remaining stable in the presence of low concentrations of free H_2S in the pore waters of the Ancient Black Sea horizon. This is quite consistent with the negligible and variable concentrations (0-20 mg/liter [7]) of free H_2S observed at present in the pore waters of the Ancient Black Sea and Neoeuxinic sediments, which may be considered to be in equilibrium with a range of sulfide and nonsulfide phases, or may result from an extremely slow process of H_2S converting to sulfides. Nonsulfide (aluminosilicate) phases are at the same time in the process of structural ordering. In relation to this it should be noted that despite a detailed study of the sediments, no crystalline forms of iron-bearing aluminosilicates (leptochochlorites) have yet been observed. This may be regarded as inconsistent with the existing model, but it can be explained logically in terms of the system outlined here. It is proposed that the ageing process of the phases described has not yet reached the stage of leptochochlorite formation, or is not yet sufficiently well-crystallized to be detected.

These facts are fully consistent with the conclusions reached in the discussion on the Neoeuxinic sediments, and allows the hypothesis which considers the Ancient Black Sea horizon to be a major source of hydrogen sulfide to be dismissed. An alternative interpretation of the data may be reached on the basis of the following known facts.

The process of microbiological sulfate reduction within marine sediments is limited by the availability of organic compounds capable of being assimilated by sulfate-reducing microorganisms (OM_{ass}) within the zone in which it occurs. These compounds are represented mainly by oxidized forms: alcohols, aliphatic acids, aldehydes, and ketones [1, 3, 7]. The formation or replenishment of these compounds occurs mainly by microbiological processes during aerobic oxidation-reduction transformations of organic detritus, but does not occur in a fully anaerobic environment.

The formation of iron sulfides in the sulfate-reducing zone may be limited either by H_2S productivity, or by the iron species available to H_2S . Where the sediment contains only a small quantity of organic matter and the surface layer is normally aerated, the first factor operates and, on account of the approximate proportionality between total and available organic matter, there is a known correlation between residual C_{org} ($\leq 2\%$) and total sulfur occurring as H_2S [7]. In this case, available forms of organic matter are apparently reproduced as a result of the influence of dissolved oxygen in the sulfate-reducing microzones. This may be confirmed by the exclusive formation of pyrite, the production of which requires the presence of an oxidizing agent.

Where there is an increased supply of organic matter, the forms of iron available to the H_2S may be exhausted. This leads, due to the occurrence of free H_2S , to the generation of extremely anaerobic conditions, which prevents the formation of OM_{ass} . As a result, once the available organic matter has been exhausted, sulfate reduction ceases. The degree of sulfide formation is therefore limited by the available forms of iron which are assessable to H_2S at the concentrations attained, which in turn depends on the reserves of OM_{ass} . The content of these forms of iron therefore strictly limits the development of processes of microbiological sulfate reduction. This inherent brake on sulfate reduction by a reverse-feedback mechanism is a typical example of the functioning of a natural biochemical system, and must apparently operate in whatever environment and at whatever scale these processes take place, from individual microzones to entire stagnant basins.

Therefore, sulfidization is ultimately limited by the content of organic matter available for microbiological sulfate reduction. The uniformly moderate level of sulfide mineralization developed within the Ancient Black Sea sediments against a background of extremely variable residual organic matter content (Figure 1) can therefore be logically explained.

Judging from the approximately uniform content of sulfides within the sediments at various depth zones within the sea the absolute quantity of OM_{ass} available to the sediments of each of these zones was also relatively constant and did not depend on the total supply of organic matter. This phenomenon would exist where there was an anomalously high level of bioproductivity in the surface waters during the Ancient Black Sea period, and this is quite consistent with the view that OM_{ass} forms during aerobic microbiological activity on organic detritus. With abundant total organic matter, the potential for oxidization in permanently aerated marine waters becomes a limiting factor in the formation of OM_{ass} . In such a regime there is therefore a relatively constant supply of OM_{ass} to the zone of sulfate reduction. The more uniform aerial distribution of OM_{ass} across the sea bed is also conducive to its increased persistence, due to the dispersion of organic detritus during microbiological activity. The supply of variable quantities of aerobically reworked organic matter to sediments mainly in deep-water areas has not been mentioned. This remains inert during subsequent anaerobic sulfate-reduction processes. In this case, as was shown above, the supply of OM_{ass} was generally insufficient to create a high enough concentration of H_2S to ensure that all the reactive iron species were sulfidized. Therefore, in sulfate-reducing zones, the potential usually remained for Fe^{2+} to combine with silicates, and only episodic increases in H_2S productivity led to sulfidization of all of the reactive iron. However, it is apparent that even in these isolated cases, the diffusive flow of H_2S occurred only locally and to a negligible extent.

It is therefore necessary fundamentally to revise not only the diffusionary model for the formation of the Neoeuxinic sulfides, but also for the mechanism by which hydrogen sulfide enters the marine water column. The data obtained show that the degree of sulfidization of total reactive Fe in the modern pelagic deposits ($>90\%$ of total reactive Fe; see Part 1) is markedly higher than in the underlying Ancient Black Sea sediments (80-85%). That is, the concentration of free H_2S within the pore fluids of the Ancient Black Sea horizon during its accumulation was on average lower than that observed now in the modern sediments (12-20 mg/liter [7]). Since sulfate reduction ceases in highly anaerobic conditions, the role of the Ancient Black Sea sediments themselves in creating the H_2S contamination of the 2-km water column is extremely insignificant.

This conclusion leads to the following alternative model for the H_2S -content of the sea.

The exceptionally high bioproductivity of the surface waters during the Ancient Black Sea period, and the corresponding precipitation of organic matter through the water column, led to the creation of a deficit of dissolved oxygen, and localization of zones of aerobic microbiological reworking of the organic detritus, primarily within the basin waters close to the surface. Therefore, the conditions necessary for sulfate reduction (a supply of OM_{ass} and the formation of anaerobic micro- and macro-zones) were established even during the process of precipitation, once the organic matter left the zone of bioproductivity. Therefore, sulfate reduction and sulfide formation began to operate immediately on the suspended organic and iron-bearing material. These processes were governed by the principles and mutual influences described above. As a result, after the reactive

forms of iron available to the H_2S were exhausted, the hydrogen sulfide in the water restricted the active reproduction of OM_{ass} and led to the low concentrations of H_2S (up to 10 mg/liter) observed in the hydrogen sulfide zone. It is clear that in the absence of any mechanism restricting sulfate reduction, the concentrations of H_2S in this zone would be considerably higher than observed, and it would have continued to precipitate intensely even at the present day.

One physical characteristic of the proposed process should be noted. According to the existing model, the waters near the sea bed were salinized, with the development of a two-layered density stratification, before the development of the Ancient Black Sea bioproductivity. Therefore, the gradually rising density of the bed invariably acted as a physical barrier to the process of sedimentation, concentrating suspended material and at the same time separating two hydrologically isolated zones. Later, during the period of intense glacial water flow, these zones must also have had different levels of aeration, which would have had an immediate influence on the nature of oxidation–reduction transformations in the suspended material, including sulfide formation.

This can be confirmed convincingly by a number of analyses on sample material, and especially by observations in the Ancient Black Sea muds in which there are correlations between the contents of total Fe, total reactive Fe, and iron sulfides.

As was noted in Part 1, the direct relationship between total Fe and total reactive Fe is most clearly marked in the deep-water deposits in the western part of the sea; in the pelagic zone in the east the relationship is much weaker, and it is absent in the upper slope (Fig. 1). This relationship, because of the independence of the aerial distribution of detrital forms of iron on the sea floor and organic matter, cannot be the result of oxidation–reduction reactions between these forms during sediment diagenesis. Therefore it must reflect a relatively uniform sedimentary supply of detrital iron in the deep-water pelagic area in comparison with a more variable supply of reactive forms. Diagenetic reactions of detrital Fe are implied by the weak or absent correlation, together with a nonuniform supply. In general this signifies the spatial subdivision of detrital and reactive forms during the process of sedimentation. For example, in slope and base-of-slope sediments, an abrupt decrease in total Fe with a negligible change in total reactive iron is observed with the onset of sapropel deposition, whereas in central parts of Khalistaz there is an increase in total Fe on account of more total reactive Fe, with similar contents of detrital forms (Fig. 1). Similar effects are caused by a very rapid intensification in microbiological oxidation–reduction reactions in detrital iron, related to the onset of processes of bioproductivity, and with transport providing more persistent forms of reactive Fe (highly dispersed, amorphous, hydrated, and combined with organic matter) to the pelagic zone.

The positive relationship between Fe within pyrite and total reactive Fe also favors the primary formation of FeS_2 within suspended material. In its turn, this relationship reflects a direct dependency between the supply to the sediment of reactive forms of iron, and organic matter capable of microbiological assimilation (OM_{ass}) which, as shown above, limits the formation of pyrite. In other words, the relationship demonstrates that these two components are supplied together to the sediment. This process, in our view, is able to take place mainly on account of the aggregation of dispersed suspended particles of OM_{ass} and reactive iron, which apparently result inevitably in microbiological oxidation–reduction transformations of detrital iron, and affect a significant proportion of the diagenetically active suspended sedimentary material.

The main reason for the formation of this type of organomineral aggregate is quite well known. They are of primary importance to certain microorganisms which adhere to the surface of suspended particles, including crystalline forms of iron, during their life cycle [11, 15]. This leads to a repeated cycle of oxidation–reduction reactions, $\text{Fe}^{3+} \rightleftharpoons \text{Fe}^{2+}$, which leads to the destruction of the crystal lattice of the iron minerals and forms highly dispersed particles of amorphous hydroxides with varying ratios of Fe(III) and Fe(II) depending on the conditions of aeration. These particles also carry colonies of microorganisms, which impregnate the residue with products of their metabolism, and which participate in the process of bioproductivity. Furthermore, they possess a high surface activity in terms of oxidized OM_{ass} (i.e. they are diagenetically active), caused by opposite surface charges in the dispersed phases, leading to varying oxidation–reduction properties. The organo-mineral aggregates are able, therefore, to play the role of microzones of sulfate reduction dispersed within the water, and to enable sulfides to form during the process of deposition.

The distribution of different iron sulfides within the total reactive Fe also favors the sedimentary origin of the main mass of the Ancient Black Sea sulfides.

The distinct predominance of iron within pyrite compared with the total nonsulfide Fe(II) and Fe within hydrotroilite should especially be noted (Fig. 1). This fact creates a doubt over the possibility of the FeS_2 having formed in the sediment pile, for the following reason.

From the beginning of the process of bioproductivity in Ancient Black Sea times, and the supply of anomalously high quantities of organic matter, the sediments were inevitably isolated from the influences of oxidizers (O^2 , S^0 and its derivatives), which may lead to the oxidation of sulfides to form pyrite: $2S^{2-} - 2e^- \rightarrow S_2^{2-}$. In these conditions, the only possible method of forming FeS_2 is the sulfidization of $Fe(III)$ compounds, which must be accompanied by the release of equivalent quantities: of Fe^{2+} , with a deficiency of H_2S ; and FeS if it is abundant ($2Fe^{3+} + 2S^{2-} = FeS_2 + Fe^{2+}$, or $2Fe^{3+} + 3S^{2-} = FeS_2 + FeS$). A deficit in the sum of these forms in comparison with the available iron within pyrite presupposes the evacuation of a significant proportion of the Fe^{2+} from the reaction zone, to the contact zones between aerated and nonoxygenated waters; i.e. in the upper parts of the water column.

As has been noted in Part 1, an increased level of sulfidization of the total reactive Fe in individual beds occurs, as a rule, due to hydrotroilite (Fig. 1). This, in all probability, reflects the continuation of sulfate reduction and sulfide formation in an anaerobic environment by the consumption of the minor remaining reserves of OM_{ass} , both during the process of sedimentation and in the upper layers of the sediment.

MODERN DEPOSITS

The account given above may be fully applicable to the late Black Sea (Recent) period of sedimentation, since the features discussed regarding the distribution of the various components are found even more distinctly.

As was shown in Part 1, the transition from the Ancient Black Sea sedimentation to the modern was accompanied within the deep-water areas of the sea by a decrease in the C_{org} content and an increase in $CaCO_3$, a marked increase in total reactive Fe within the total Fe, and also an increase in its level of sulfidization on account of pyrite, with a simultaneous reduction in the hydrotroilite composition (Fig. 1). The generally increased content of sulfides compared to the Ancient Black Sea sediments obviously contradicts the existing model, in which the end of sapropel deposition would lead to a fall in the level of sulfate reduction, and which does not consider the modern sediments as such a major source of H_2S [14]. A satisfactory explanation for the observed sulfide distribution can be provided by the hypothesis of sulfide formation from suspended material.

The end of the Ancient Black Sea period of deposition coincided with an abrupt reduction in the deposition of organic muds. This occurred, on the one hand, as the result of a reduced bioproductivity of the surface waters and, on the other hand, due to the major supply of $CaCO_3$ to the sediments. One of the reasons for the fall in bioproductivity may have been the exhaustion of biogenic elements due to the complete displacement of the Neoeuxinic waters, i.e. the approach of the high-density bed of the stratified water column to the bioproductive photic zone. Partial mixing of the high-density bed with the zone of active aeration may have led to a substantial increase in the degree of aerobic microbiological activity on the suspended organic material and detrital iron concentrated within this bed. This may have led to an increase in total reactive Fe within the total Fe, and optimum conditions for pyrite formation. The improved aeration of the organic matter (and of the residence time within the aerated zone) would lead not only to an increase in reserves of OM_{ass} , i.e. a general growth of iron sulfides within the total reactive iron, but also to an increase in the proportion of OM_{ass} expended in the process of FeS_2 formation. With a limited productivity of OM_{ass} , this would correspond to a decrease in the proportion which could in future be expended in strongly anaerobic conditions of sedimentation and diagenesis, and therefore would reduce the hydrotroilite composition. With the high-density bed close to its modern position, the major effects are noted within deep-water sediments and are usually absent in the deposits of the shelf and the upper slope.

Therefore it may be proposed that with the end of the Ancient Black Sea period, despite the decrease in bioproductivity, the processes of sulfate reduction and sulfide formation not only did not decrease, but they increase somewhat due to more thorough reworking of organic matter into an oxidized, diagenetically active form capable of sulfate reduction. The regime which developed then has apparently remained in all its main features until the present day.

The possibility of the processes described operating extensively on suspended material within the water column of the modern sea has already been noted [10, 11]. In particular, microbiological studies have shown the existence of a zone of active sulfate reduction within an area of greatly increased density. This zone, in terms of the numbers of bacteria and the rate of H_2S formation, is comparable with the zone found on the surface of the deep-water sediments [11]. Direct confirmation of sulfide formation during the process of sedimentation is available from the occurrence within the suspended material in the Black Sea of framboidal pyrite [17]. At the same time, sulfate reduction within the modern deep-water sediments is highly restricted. The number of sulfate-reducing bacteria within the first few centimeters of the sediment pile falls sharply, and below a depth of 5 cm the modern sediments are quite rich in organic matter (C_{org} 3-5%), but are microbiologically sterile [11]. This fact demonstrates

convincingly that sulfate reduction ceases in anaerobic conditions once the forms of organic matter suitable to support microbiological organisms are exhausted.

The ceasing of sulfate reduction within the surface layers is in accordance with the similarity between the concentrations of free H_2S in the pore waters of pelagic sediments (8-12 mg/liter) and those in the water column (~10 mg/liter). The virtual absence of any H_2S concentration gradient would permit only a negligible diffusive flow of H_2S from the sediment surface into the overlying waters. An increase in diffusive flow may occur within the sediments on the upper part of the slope, where an increased concentration of free H_2S have been observed in the surface beds (12-20 mg/liter [7]). The tendency of these zones to be restricted to basin slopes is apparently a reflection of the distribution of diagenetically active organic matter, and its increased concentration in these areas. These slight variations in the intensity of H_2S formation do not alter the general conclusion reached for the deep-water sediments that sulfate reduction stops within the surface beds (0-5 cm), and that its potential as an H_2S source is highly limited.

* * *

The main conclusions arising from this analysis of the distribution of reactive forms of iron within the Late Quaternary sediments of the Black Sea can be summarized as follows:

1. The distribution of Fe and S species within the Neoeuxinic deposits contradicts the theory of a diffusive mechanism for the formation of sulfide mineralization, but is consistent with sulfate reduction and sulfide formation within the Neoeuxinic basin itself.

2. The role of the Ancient Black Sea organic muds as a source of H_2S by diffusion into the underlying sediments and the overlying water column is limited, and insufficient as an explanation for the formation of the Neoeuxinic sulfides, and the hydrogen sulfide content of the water column.

3. The proportion of reactive and detrital forms of iron observed within the Ancient Black Sea and modern sediments is the result of several stages of microbiological oxidation-reduction reactions, mainly on suspended iron-bearing sedimentary material.

4. During the aerobic reworking of organic matter within the zone of bioproductivity, reserves of oxidized, diagenetically active forms of organic matter were created, the size of which limited the process of sulfate reduction. Where the total amount of organic matter was abundant, these reserves were limited by the oxidizing potential of the permanently aerated regime of the surface waters, and therefore were relatively constant over different topographic zones and geographical regions of the sea.

5. During the aerobic reworking of organic matter, reactive forms of iron formed aggregates with diagenetically active organic matter, resulting in sulfate reduction and sulfide formation taking place in suspension.

6. The proportions of the various chemical forms of sulfides are mainly determined by the oxidation regime within the zone of sulfide formation. Limited (optimal) aeration produces FeS_2 , and highly anaerobic conditions produce acid-soluble sulfides. The greatest amount of FeS_2 forms in the contact zone between the aerated and oxygen-free waters.

7. The exhaustion of reserves of water-soluble oxygen and reactive iron accessible to H_2S led to the onset of hydrogen sulfide contamination of the water column. Only the final stage of this process takes place within the sediment itself.

This interpretation of the processes of formation of reactive forms of iron differs substantially from the earlier theory of the geochemical evolution of the Black Sea during the Late Quaternary, and therefore certainly needs to be confirmed and amended on the basis of a range of lithological, microbiological and chemical-mineralogical studies.

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THE BEHAVIOR OF TRANSITIONAL METALS IN THE PORE WATERS OF BALTIC SEA SEDIMENTS

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The distribution of transitional metals within pore waters of sediments in different parts of the Baltic Sea is discussed on the basis of data in the literature. It is shown that the content of dissolved manganese and to a lesser degree of zinc correlates with their content in mobile form within the solid phase. For Fe, Cu, and Ni, there is no such relationship, but a correlation is observed between dissolved Zn, Ni, and Fe. The ratio of the metal content of pore and bottom waters varies from 1 to 115 (Mn), which provides suitable conditions for their diffusion out of the sediments.

The behavior of metals in the fluid phase of modern sediments is one of the most objective indicators of their geochemical activity during early diagenesis. The results of relevant studies are used here to reconstruct the conditions of formation of a number of types of sedimentary ore. The results are important for solving ecological problems related to the man-made efflux of metals into the water body.

A number of papers have considered the problem of the behavior of metals in pore waters of deep and shallow sediments [1, 3, 5, 8-12, 14-18, 20-23, 25-30, 33-38, etc.]. Their main interest in terms of the Baltic Sea has been with regard to changes in the original hydrochemical regime, the increasing concentrations of a number of heavy metals within the sediments, and also in relation to the increasing anthropogenic influences.

The salinity of the Baltic Sea pore waters has been described in [5, 8-10, and 17], and the distribution of microelements in [5, 16, and 25]. The most representative data were obtained during trip 26A of the research ship *Akademik Kurchatov* (1978), when a wide range of studies of both the solid and fluid phases of the sediments were completed. The bulk content of mobile forms of transitional metals was determined within the solid phase, including manganese, iron, copper, zinc and nickel. A total of 82 pore-water samples were studied, 39 of which were taken from the Gotland Basin (18 from bottom-samples and 21 from cores) and the remainder from the Gulf of Riga and the Klaipeda region (Fig. 1). Furthermore, 20 metal analyses were undertaken on samples of bottom water.

The results of these analyses have been published in [5, 16], although the unique material obtained was only partially systematized. The aim of this paper is to provide these data in a more fully systematized and interpreted form so that similar studies may be continued, extended, and completed.

A general description of the depositional environment, and a number of aspects of the diagenesis of sediments in the Gotland Basin and the Gulf of Riga are provided in [5, 7, and 13], and other characteristics of the behavior of a number of elements during these processes are discussed in [24, 26, 31, 32, and 38].

According to the available data, the water column in the Gotland Basin consists of three layers, separated by oxidation-reduction geochemical barriers. The upper layer is relatively fresh and oxidizing (0-60 m), and is underlain first by the intermediate layer (60-170 m), and then by the basal layer, which is saline and stagnant (170-246 m). A salinity of 12-13‰ is typical of the basal layer; it has a constant temperature of about +4°; during periods of stagnation there is an absence of free oxygen, the occurrence of hydrogen sulfide, and an enrichment in biogenic elements and a number of metals including in particular manganese and iron. At the margin of the intermediate and basal layers, at a depth of 140-160 m, a more turbid layer is noted, containing around 2.5 mg/liter of suspended material. This is composed primarily of coagulates of manganese oxide, forming at the contact of the oxygenated and H₂S-bearing waters.

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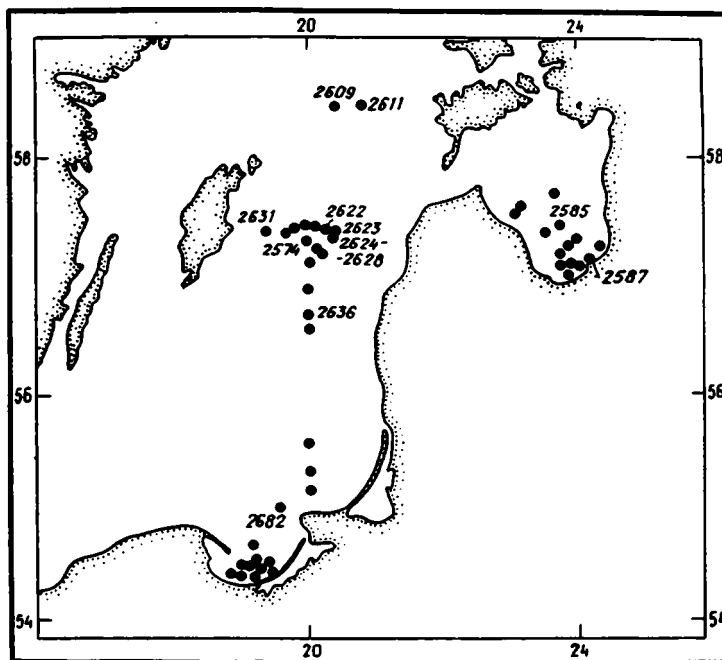


Fig. 1. Stations at which pore-water samples were taken during trip 26A of research ship *Akademik Kurchatov*.

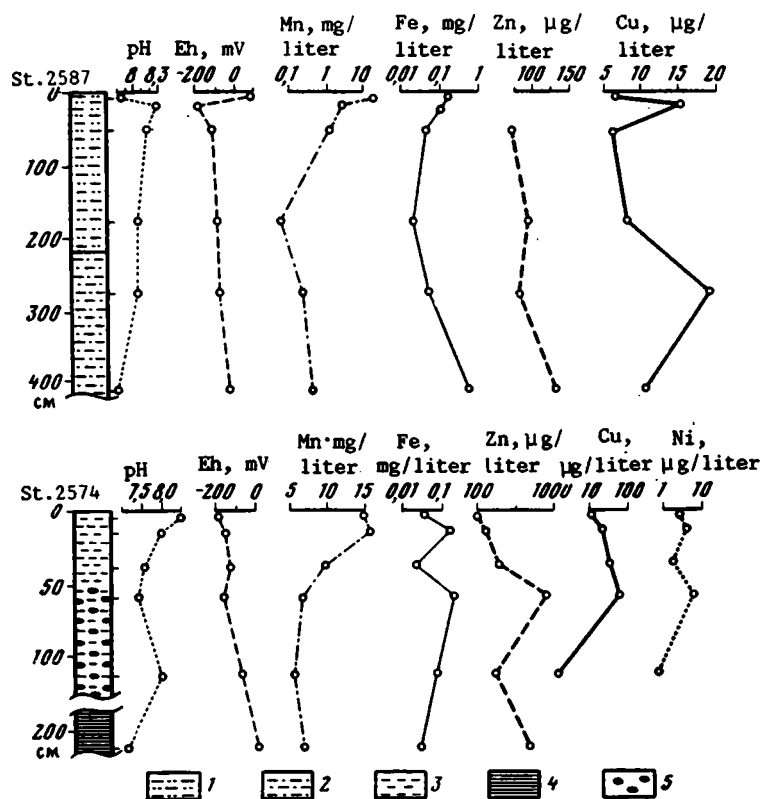


Fig. 2. Metal distribution in pore waters of sediments in the Gotland Basin: 1) fine-silty mud; 2) silty-pelitic mud; 3) pelitic mud; 4) homogeneous clay; 5) hydrotroilite.

TABLE 1. Metals within Sediments, Pore and Bottom Waters from Bottom Samples within the Gotland Basin [5, 16]

Station number	Depth, m	Sediment horizon, cm	pH	Eh, mV	C _{org}	Mn			
						I	II	III	IV
2622	240	4-18	7.1-7.6	-230÷-220	3.06-3.29	0.21-0.32	0.08-0.09	17.3-18.2	0.15
2623	220	0-25	7.3-7.5	-230÷-220	3.15-3.23	0.08-0.25	0.02-0.07	7.0-7.1	0.8
2624	205	0-25	7.4-7.6	-230÷-220	3.37-3.56	0.04-0.05	0.03	3.1-3.7	-
2625	170	0-22	7.3-7.5	-240÷-215	3.33-3.64	0.04-0.05	0.02-0.04	0.69-0.92	1.0
2626	127	0-23	6.9-7.1	-190÷-50	2.12	0.04-0.05	0.01	0.18-0.44	-
2627	120	0-23	6.9-7.0	-180÷-170	3.00-3.17	0.02-0.03	0.01-0.02	0.18	-
2628	90	0-21	-	+20÷+30	3.27-3.59	0.02-0.03	0.01-0.02	0.05-0.20	0.05
Fe									
2622	240	4-18	7.1-7.6	-230÷-220	3.06-3.29	5.2-5.7	1.25	0.17-1.1	0.09
2623	220	0-25	7.3-7.5	-230÷-220	3.15-3.23	5.6-6.3	1.89-2.29	0.35-1.65	0.04
2624	205	0-25	7.4-7.6	-230÷-220	3.37-3.56	5.3-5.8	1.73-2.16	0.3-0.9	-
2625	170	0-22	7.3-7.5	-240÷-215	3.33-3.64	4.4-5.0	1.29-1.57	0.05-0.21	0.02
2626	127	0-23	6.9-7.1	-190÷-50	2.12	4.4-5.0	1.28-1.75	0.21	-
2627	120	0-23	6.9-7.0	-180÷-170	3.00-3.17	4.0-4.8	0.94-1.05	0.16-0.33	-
2628	90	0-21	-	+20÷+30	3.27-3.59	3.8-3.9	1.18-1.21	1.0-1.8	0.02
Zn									
2622	240	4-18	7.1-7.6	-230÷-220	3.06-3.29	165-180	67-81	10-50	25
2623	220	0-25	7.3-7.5	-230÷-220	3.15-3.23	160-190	64-82	115-155	20
2624	205	0-25	7.4-7.6	-230÷-220	3.37-3.56	149-196	59-71	60-75	-
2625	170	0-22	7.3-7.5	-240÷-215	3.33-3.64	150-173	53-83	45-140	80
2626	127	0-23	6.9-7.1	-190÷-50	2.12	150-173	47-69	45-1270	-
2627	120	0-23	6.9-7.0	-180÷-170	3.00-3.17	132-154	58-88	290-660	-
2628	90	0-21	-	+20÷+30	3.27-3.59	101-160	41-89	150-10000	10
Cu									
2622	240	4-18	7.1-7.6	-230÷-220	3.06-3.29	51-53	32-35	8-37	18
2623	220	0-25	7.3-7.5	-230÷-220	3.15-3.23	52-54	35-38	18-20	13
2624	205	0-25	7.4-7.6	-230÷-220	3.37-3.56	50-53	33-36	27-31	-
2625	170	0-22	7.3-7.5	-240÷-215	3.33-3.64	41-45	30-24	22-43	37
2626	127	0-23	6.9-7.1	-190÷-50	2.12	37-38	22-26	21-26	-
2627	120	0-23	6.9-7.0	-180÷-170	3.00-3.17	36-45	24-32	21-35	-
2628	90	0-21	-	+20÷+30	3.27-3.59	32-37	19-26	110-520	22
Ni									
2622	240	4-18	7.1-7.6	-230÷-220	3.06-3.29	75-78	25-26	8-29	16
2623	220	0-25	7.3-7.5	-230÷-220	3.15-3.23	75-80	25-35	5-8	9
2624	205	0-25	7.4-7.6	-230÷-220	3.37-3.56	60-70	23-28	5	-
2625	170	0-22	7.3-7.5	-240÷-215	3.33-3.64	50-56	18-24	7	7
2626	127	0-23	6.9-7.1	-190÷-50	2.12	50-77	17-19	7	-
2627	120	0-23	6.9-7.0	-180÷-170	3.00-3.17	51-62	14-18	6-7	-
2628	90	0-21	-	+20÷+30	3.27-3.59	35-45	19	0.5-0.33	6

Note: I) Total content of element in dry sediment, %; II) content of mobile form of element within sediment (Mn, Fe in %, others in 10⁻⁴%); III, IV) metal content in pore and bottom waters respectively (Mn, Fe in mg/liter, others in µg/liter).

The thickness of Holocene sediments in the Gotland Basin is from 40-80 cm in the peripheral areas, to 223-367 cm in the central zone. This corresponds to a sedimentation rate of from 6-12 to 30-50 cm/thousand years. Sediments in the upper layer are represented by terrigenous pelitic muds containing up to 4% C_{org}. Within the permanently stagnant zone the muds are black and greenish-black, whereas in the margins of the basin they are greenish-black, greenish-grey, and often banded. The first two types are highly reduced (Eh from -215 to -240 mV), with a water content of 72-76%, and a pelitic fraction of 80-90%. Transitional sediments on the eastern slope of the basin are silty pelites and fine silty muds, with Eh values of -50 to +80 mV, and a water content of 65-70%.

In the Gulf of Riga, which according to the classification of Emel'yanov [7] is a shallow-water quartz-feldspar-silicate region, the sediments are coarse-grained and oxidized. The upper layer is enriched in Mn, Ni, and Co, and shotlike concretions are found on the surface.

TABLE 2. Relative Composition of the Mobile Forms of Metals (% of total) in Sediments of the Gotland Basin

Station number	Mn	Fe	Zn	Cu	Ni
2622	33	24	42	64	33
2623	30	36	43	69	36
2624	70	34	38	67	40
2625	65	30	44	72	41
2626	22	33	36	66	30
2627	55	26	50	70	28
2628	58	31	48	67	48
Average	48	30	43	68	37

Table 1 is a generalized compilation of the analytical results of the material discussed, and provides the metal content of the solid and fluid phases for the sediments in the Gotland Basin obtained from bottom samples. Figure 2 illustrates the distribution of metals within the pore water of cores from the same basin.

Data on the form of occurrence of metals in the solid phase of the sediments show that the proportion of the mobile form comprises on average: for copper, 68%; manganese, 48%; zinc, 43%; nickel, 37%; and iron, 30% (Table 2).

Apparently the microelement composition of the pore-waters depends on this reserve of mobile metals.

It is notable that in the sediments of stations 2622-2627 (Table 2) the physicochemical conditions are reducing (Eh from -240 to -170 mV), with the exception of just one horizon in Station 2626 (-50 mV), and in the sediments from Station 2628 the conditions are transitional from reducing to oxidizing (Eh from +20 to +30 mV). For iron and copper the relative contents of mobile forms in these and other sediments are practically identical, whereas for manganese, zinc, and nickel they are higher in the transitional sediments than in the reducing sediments. Apparently, with the strengthening of the reducing processes, the "elevated" quantities of metals may pass into the pore water and their relative proportion in the mobile form of the solid phase falls.

The average content of metals within the pore waters is given in Table 3, separated into values for the Gulf of Riga, and bottom samples and core samples from the Gotland Basin. In each of these groups the samples are divided into reducing and transitional sediments.

This comparison shows that in the whole succession of sediments within the shallow-water Gulf of Riga, and in the upper horizons of the deep-water sediments of the Gotland Basin, manganese is actively accumulating within the fluid phase of the reduced sediments, but significantly less intensely within the transitional sediments. In the sediment succession in the Gotland Basin the picture is different, since the manganese is distributed uniformly within the pore waters, and is independent of the physicochemical conditions. This may be related to its diffusion out of the reduced and into the transitional sediments, or with the formation in the latter of stable metal-organic complexes, which are proposed on the basis of a correlation which is sometimes observed between the content of dissolved manganese and organic carbon [22, 26].

In contrast to manganese, iron within the pore waters of the sediments in the Gulf of Riga does not react to changes in the oxidation-reduction conditions, and accumulates in equal amounts within the pore waters both of the reduced and the transitional sediments. At the same time, the concentration of iron within the surface layer and in the sediment succession in the Gotland Basin is 2-3 times higher in the pore waters of the transitional sediments compared to the reducing sediments. A similar pattern is noted for nickel.

A characteristic of the behavior of zinc and nickel is displayed in a number of horizons of the upper layer of the transitional sediments in Gotland Bay, where their concentrations rise sharply. Elsewhere, the distribution of zinc in the pore waters is similar to that of manganese.

Therefore, in comparing the average metal contents within the range of pore waters discussed, they can quite clearly be subdivided into two groups: Mn-Zn and Fe-Cu-Ni. A comparison of the concentrations of mobile forms of metals in the sediments and dissolved metals within the pore waters reveals a trend towards a direct correlation for manganese (Fig. 3), a rather poorer correlation for zinc, and no correlation for iron, copper and nickel. It was noted earlier that generally for marine and oceanic sediments there is a negative correlation between the content of manganese in solution and in the solid phase (including both the total content and the mobile forms). Pelagic clays enriched in manganese are characterized by a minimal content of dissolved manganese, whereas in coastal and hemipelagic sediments the reverse is often observed [1]. This case, apparently, is an example of a locally applicable correlation with a relatively low manganese content in the solid phase and high

TABLE 3. Average Content of Metals in Pore Waters Calculated from Data in [5, 16]

Region	Water depth, m	Mn		Fe		Zn		Cu		Ni	
		I	II	I	II	I	II	I	II	I	II
Gulf of Riga (cores)	30-42	3,06	0,23	0,39	0,39	59	142	16	15	2,5	< 3
Gotland Basin (bottom samples)	90-240	4,9	0,12	0,47	1,4	243	5000	26	315	9	16
Gotland Basin (cores)	130-229	4,7	4,2	0,34	0,6	236	253	34	46	7,5	15
Average		4,4	1,5	0,4	0,8	180	1800	25	125	6	11

Note. I) Reduced sediments; II) sediments of transitional type. Contents of Mn and Fe in mg/liter, other elements in $\mu\text{g/liters}$.

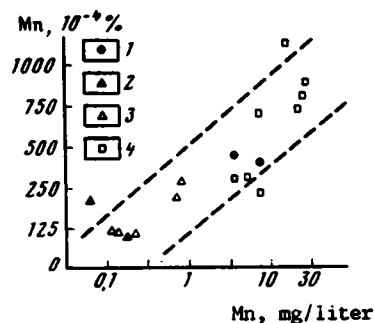


Fig. 3. Relationship between the manganese content in the pore waters and the content of its mobile form within the sediments. 1) Gulf of Riga, 2) transitional zone, 3, 4) marginal and deep-water parts of the Gotland Basin respectively.

in the fluid. Sediments with an increased manganese content (up to over one percent) are found within basins of the Baltic Sea [7, 13], but unfortunately were not present in the samples studied.

The ratio between the concentration of metals within the pore waters of the upper horizons of the sediments and the bottom waters in contact with them varies widely, particularly for manganese and iron (Table 4). The maximum value of this ratio (115) was established for manganese within the deep-water zone of the Gotland Basin, where free hydrogen sulfide is present (0.26 ml/liter according to data from trip 26A of the research ship *Akademik Kurchatov*). In the absence of an isolating oxidizing film on the sediment surface, the manganese might have been free to diffuse into the bottom water, and its concentration above and below the sediment-water interface would have become equal. Probably in this case this did not happen, since the manganese within the pore water is in the form of metallic-organic complexes, and does not react at such a concentration gradient. On the other hand, however, at Station 2625, on the margin of the Gotland Basin, the manganese content (and that of other metals) in the pore and bottom waters is virtually identical, which demonstrates the free exchange of metals through the sediment-water interface. Within sediments of transitional type, the concentration gradient for manganese within the pore and bottom waters again grows, particularly at the shallowest Station 2585 (Gulf of Riga). Here, the C_{org} content in the sediments in a profile through the Gotland Basin remains generally at one level, within the range 3.00-3.59% (Table 1). Therefore, a proper study would require the relationship of the manganese content within the sediment pore waters of the Gotland Basin with the depth, the physicochemical environment, and other factors, to be investigated further.

The ratio between the contents of other metals within the pore and bottom waters varies irregularly, from values of 15 (zinc) to 70 (iron). However, they all approach a value of one on the margin of the Gotland Basin (Station 2625), and all, with the exception of nickel, display a sharp peak on the border of the transitional sediments (Station 2628). The latter is probably due to the formation of an oxidized film on the surface of the reduced sediments, preventing exchange through the sediment-water interface.

TABLE 4. Ratios between the Contents of Metals in the Pore Waters of the Upper Layer of the Sediments and the Bottom Waters

Station number	Depth, m	Conditions within sediment		Mn	Fe	Zn	Cu	Ni
		pH	Eh					
2622	240	7,1 ¹	-220(H ₂ S)	115	12	2	2	2
2574	220	7,7	-200(H ₂ S)	43	0,7	4	0,6	0,5
2623	220	7,3 ²	-220	9	9	8	1,4	0,8
2625	170	7,5	-215	0,9	1	1,7	1,2	1
2609	140	7,6	-100	-	14	5,2	2,8	1
2628	90	- ³	+30	4	50	15	24	0,1
2585	42	7,5	-50	56	0,5	1	1	-

Note. pH within bottom waters: 17.1; 27.2; 37.3.

TABLE 5. Mn/Fe and Zn/Cu Ratios within Sediments, Pore and Bottom Waters

Station number	Depth, m	Mn/Fe				Zn/Cu			
		I	II	III	IV	I	II	III	IV
2622	240	0,05	0,06	61	1,7	3,3	2,2	1,3	1,4
2623	220	0,03	0,02	12	20	3,2	2,0	7,0	1,5
2624	205	0,009	0,014	7,7	-	3,3	1,9	2,3	-
2625	170	0,01	0,02	9,2	5	3,6	2,1	1,6	2,2
Average for deep-water zone		0,025	0,029	22	9	3,3	2,0	3,0	1,7
2626	127	0,01	0,007	1,4	-	4,4	2,2	25	-
2627	120	0,005	0,01	0,8	-	3,5	2,6	20	-
2628	90	0,006	0,014	0,11	2,5	3,7	2,8	45	0,5
Average for shallow-water		0,007	0,010	0,8	2,5	3,9	2,5	30	0,5

Note. I) Total sediment sample; II) mobile forms of metals in sediment; III) pore waters; IV) bottom waters.

The pairs of ratios used to characterize geochemical processes in metals vary in the phases described in the following way. In the order: [sediment — mobile phase in the sediment — pore waters — bottom waters], the average values of Mn/Fe, according to Table 5, are: in deep water, [0.025 — 0.029 — 22 — 9], in shallow water, [0.07 — 0.01 — 0.8 — 2.5]. The values for Zn/Cu in the same order are: in deep water, [3.3 — 2 — 3 — 1.7], and in shallow water, [3.9 — 2.5 — 30 — 0.5].

It is evident that within the sediments, pore, and bottom waters of the deep-water zone the Mn/Fe values are generally higher than in shallow water. The ratio is particularly high within the pore waters of the former zone, being two orders of magnitude higher than in the sediments. In the shallow-water zone the ratio increases in comparison with the sediments by an order of magnitude, but is still less than one. This situation is obviously related to the predominance of iron over manganese within marine iron—manganese concretions, which are formed in normally aerated conditions in deposits close to the sediment—water interface as a result of the diagenetic movement of metals from the underlying sediments. In pore waters of oceanic pelagic sediments, the Mn/Fe values are usually higher than one, since oceanic diagenetic concretions have an elevated manganese content [2].

The Zn/Cu ratio is relatively uniform within sediments and waters of the deep-water zone. Within shallow water the values remain approximately the same in the sediments, but increase by an order of magnitude in the pore waters. This corresponds to a predominance of copper over nickel within iron—manganese concretions of the Baltic and other seas [4, 6, 19, 32]. High Zn/Cu values are also observed in the pore waters of oceanic pelagic sediments, whereas in oceanic iron—manganese concretions, copper predominates over zinc, probably because of the nature of the mineralogy and the lower rate of growth of these concretions.

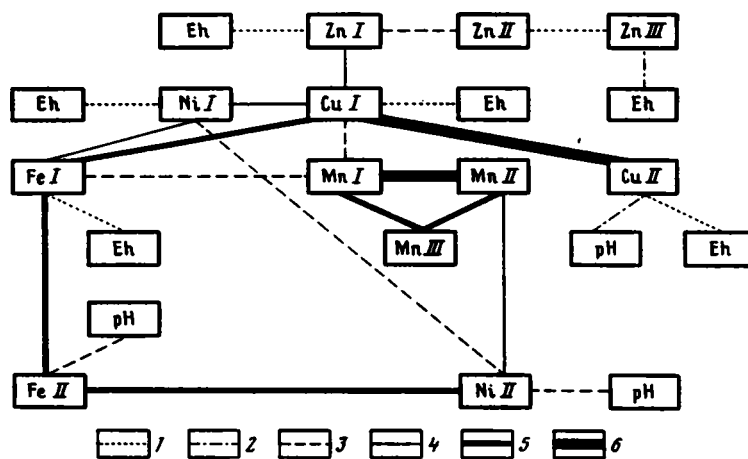


Fig. 4. Correlations between the contents of metals in sediments and in pore waters with pH and Eh. Correlation coefficients: 1) <0.5; 2) 0.5-0.6; 3) 0.6-0.7; 4) 0.7-0.8; 5) 0.8-0.9; 6) >0.9.

In the opinion of Hallberg [24], the ratio of copper and zinc can be used to reconstruct the ancient hydrochemical regime of a basin, with the period of stagnation being related to the copper enrichment in the sediment. In the case under discussion, no such difference is noted between the upper beds of sediment in the stagnant and aerated zones, but it is clearly exhibited within the pore waters.

In order to evaluate the relationships between metals in the various phases discussed, and also their relationships with the physicochemical indicators of the environment (Eh and pH) their paired correlation coefficients have been calculated. Thirteen bottom samples were used from the Gotland Basin, for which a full set of analytical data was available. The results are presented in graphic form on Fig. 4, and may be summarized as follows:

- The total content of metals in the solid phase; (nickel correlates with iron; copper correlates with iron and, to a lesser degree, with manganese, and also with nickel and zinc).
- Mobile forms of metals in the solid phase; (manganese, iron, copper, and to a lesser degree zinc and nickel, correlate with the total contents of the same metals; furthermore, nickel correlates with mobile iron and manganese).
- Metals in the fluid phase; (manganese correlates with its total content, and the content of the mobile form in the solid phase; zinc correlates with nickel, and to a lesser degree with iron).

Therefore, in the material studied the metals, except manganese, generally do not correlate in the solid and fluid phases. This is apparently due to their migration and their form of occurrence in the pore waters. Further studies of the behavior of metals during the diagenesis of sediments in the Baltic Sea should determine the nature of their occurrence simultaneously in the solid and fluid phases. This would help in the understanding of the exchange mechanism across the sediment-water interface.

In conclusion, the author expresses gratitude to Dr. H. Likke-Anderson (Arhus University, Denmark) for his help in completing the correlation calculations on a computer during the 24th trip of the research ship *Professor Shtokman* in the Baltic Sea in 1989.

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BIOGENIC COMPONENTS IN BOTTOM SEDIMENTS, EASTERN EQUATORIAL PACIFIC

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UDC 550.4:551.79(266)

Distribution patterns of biogenic components (CaCO_3 , $\text{SiO}_{2(\text{am})}$, C_{org}) and phosphorous have been determined in sediments of the eastern equatorial Pacific. It is shown that the distribution of these components is related to climatic, circumcontinental, and vertical zonation of deposition. Hydrothermal activity was not found to have an appreciable effect on pelagic biogenic sedimentation.

Biogenic sediments in the oceans consist mainly of calcium carbonate, amorphous silica, and organic matter or, more properly, organic carbon. To a lesser degree they include various shelly remains or fish scales, which are composed of calcium and P_2O_5 . We have examined the distribution of these chemical components in the eastern part of the northern half of the Pacific equatorial zone. The goals of this article are as follows: 1) to locate more precisely the calcite compensation depth (CCD), and 2) to determine the variations in the content of biogenic components ($\text{SiO}_{2(\text{am})}$, C_{org} , and CaCO_3) and P in bottom sediments.

Location and Geographic Character of the Region. The study area lies between the equator and 20°N and is thus located in the so-called equatorial facies zone of high bioproductivity [16], or the equatorial diatom-radiolarian zone [7, 8]. The region extends laterally from 150° to 90°E , that is, it includes the Guatemala basin and part of the northeastern plain (Fig. 1). The area is bounded on the north by the Clarion fracture zone, and the Clipperton fracture zone is located in the middle. The Guatemala Basin, with depths mostly in the 3600-3700 m range, is separated from the open Pacific (mostly 4500-5500 m deep) by the north end of the East Pacific Rise (EPR); to the south is the Galapagos Rise, and to the east, the Cocos Ridge and the continental slope of South America. This region is known for intense hydrothermal activity [10, 13, 15, 25]. A second feature is

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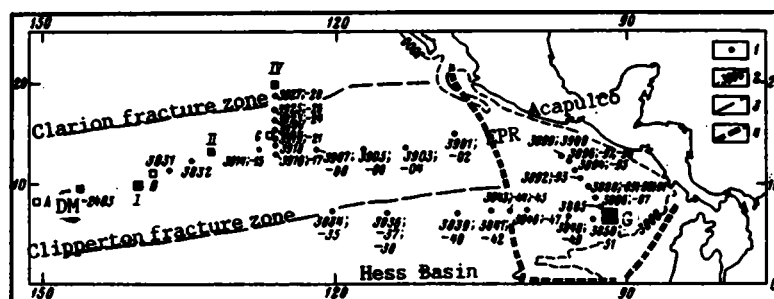


Fig. 1. Study area in the Pacific Ocean (expedition DM-41). 1) Geologic stations; 2) 3000 m isobath; 3) fault; 4) ridge or rise; I) (sta. DM-3830); II) (sta. DM-3833 and DM-3913); IV and G) (sta. from DM-3852 to 3884): traverses of expedition DM-41; sta. DM-2483, traverse of expedition DM-29; A, B, and C: traverses of the American expedition DOMES [26]; EPR: East Pacific Rise.

its very high biological productivity: east of the EPR it is usually $100-500 \text{ mg} \cdot \text{C} \cdot \text{m}^2$ per day, while to the west it is $150-250 \text{ mg} \cdot \text{C} \cdot \text{m}^2$ per day [11].

Previous Studies, Material, and Methodology. The bottom sediments of the equatorial Pacific radiolarian belt has been thoroughly studied: maps have been published of bottom sediments [4, 7], facies [8, 16], and distribution of CaCO_3 , $\text{SiO}_{2(\text{am})}$ [13], and C_{org} .

Important analytical data obtained during the 9th voyage of the research vessel *Dmitri Mendeleev* (expedition DM-9) were used by the author in compiling a map of the distribution of biogenic components in the sediments. Detailed studies in the region were completed in the following traverses: DM-2483 [8] and A, B, and C of the American expedition DOMES [26]. Such studies were virtually nonexistent in the Guatemala Basin.

Additional studies were conducted in the region, including the Guatemala Basin, in 1988 during the 41st voyage of the *Dmitri Mendeleev* (expedition DM-41, under the direction of A. G. Rozanov and with the participation of the author). Bottom sediments were sampled by a team under the direction of V. N. Stepanov, E. A. Kontar', and N. S. Skornyakova. The stratigraphy of the sediments was determined from micropaleontologic data by S. B. Kruglikova and V. V. Mukhina under the direction of M. S. Barash. The author has improved a little on the lithologic characterization of all of the samples. Chemical analyses were carried out in the laboratories of the Geology of the Atlantic, Atlantic Division, Institute of Oceanology, Academy of Sciences (Kaliningrad) by N. G. Kudryatsev, V. A. Kravtsov, T. V. Firsov, and G. S. Khandros. CO_2 and C_{org} were determined by volumetric methods on an AN-7629 analyzer, $\text{SiO}_{2(\text{am})}$ by chemical methods in sodium extract, and P by colorimetric methods. Part of the CaCO_3 determinations were carried out during the expedition by N. V. Osadcha, and part of the Fe and Mn determinations were done on cores by N. N. Zavadzka.

Type of Sediments. Pacific bottom sediments have been studied by numerous investigators, but the most complete maps have been published in [2, 3, 5, 8, 16]. The author has studied these maps and made use of the data [17], and has also used the classification accepted in the Institute of Oceanology [3, 16], in compiling the map of bottom sediment distribution (Fig. 2).

Miopelagic clays occur in the north of the region. They are brown to reddish-brown soft oozes with numerous ferromanganese concretions (see Fig. 2, 2). The CaCO_3 content in the clays is usually $<1\%$, $\text{SiO}_{2(\text{am})}$ is $1-2\%$, C_{org} is $0.10-0.25\%$, and P is $0.10-0.15\%$. The clay is slightly manganiferous ($0.37-1.47\%$ Mn), and the Fe content is $5.29-6.35\%$.

To the south, miopelagic clays grade into reddish-brown slightly siliceous pelagic clays with diatoms and radiolarians. The $\text{SiO}_{2(\text{am})}$ content in these clays, determined chemically in sodium extracts, ranges from 5.0 to 7.7% . If one considers that the chemical analyses understate (by about 1.6 times) the true content of biogenic silica, then the $\text{SiO}_{2(\text{am})}$ content in the slightly siliceous miopelagic clays actually ranges from 8.0 to 12% . The C_{org} content in these oozes is usually $0.20-0.35\%$ (in some cases $0.11-0.20\%$), the phosphorous is $0.09-0.24\%$, iron is $3.78-5.80\%$, and manganese is $0.18-1.40\%$. The slightly siliceous oozes, like the miopelagic clays, are thus slightly manganiferous. South of these clays, at depths of less than 4400 m , there are coccolith-foraminifera oozes with CaCO_3 contents of $50-72\%$. These sediments lie on the EPR. The coccolith-foraminifera oozes are

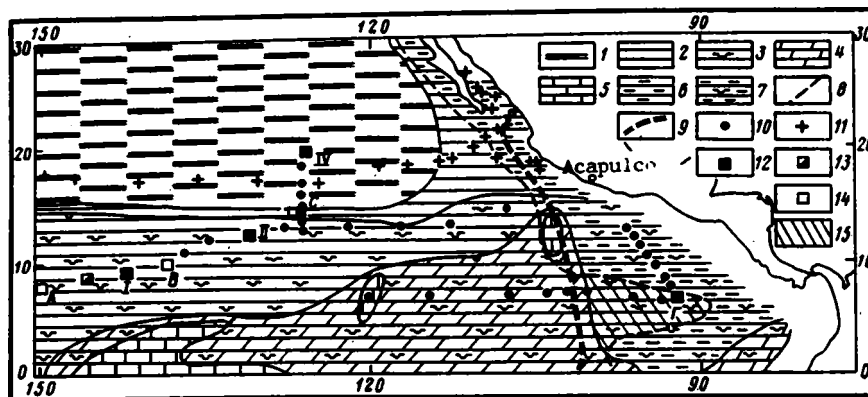


Fig. 2. Map of the distribution of bottom sediments (0-3-cm layer). 1) Miopelagic clay, brown, sometimes with zeolites, often with ferromanganese concretions ($\text{CaCO}_3 \leq 1\%$, $\text{SiO}_{2(\text{am})} \leq 1\%$, C_{org} 0.25%); 2) miopelagic clay, brown and reddish-brown, with higher content of CaCO_3 (1-10%) and $\text{SiO}_{2(\text{am})}$ (2-5% diatoms, radiolarians, and silicoflagellates), C_{org} 0.25-0.40%; 3) slightly siliceous clayey ooze (or miopelagic clay) ($\text{SiO}_{2(\text{am})}$ 5-10% or 10-15% diatoms, radiolarians, and silicoflagellates), CaCO_3 1-10%; 4) slightly siliceous, marly coccolith-foraminifera ooze (CaCO_3 50-70%, $\text{SiO}_{2(\text{am})}$ 5-10% or 10-15% diatoms, radiolarians, and silicoflagellates); 5) coccolith-foraminifera carbonate ooze (CaCO_3 70-85%); 6) terrigenous, brown hemipelagic ooze with higher C_{org} content (1-7%); 7) same, but slightly siliceous ($\text{SiO}_{2(\text{am})}$ 5-10% or 10-15% diatoms, radiolarians, and silicoflagellates), C_{org} 1-2%; 8) contact between oxidized (brown to reddish brown) and reduced (gray) oozes; 9) crestal zone of East Pacific Rise; 10, 11) stations of expeditions DM-41 and DM-9, respectively; 12-14) traverses of expeditions DM-41, DM2483, and DOMES, respectively; 15) miopelagic manganiferous (Mn 1-8%) and siliceous ($\text{SiO}_{2(\text{am})}$ 5-10%) clay with higher C_{org} content (0.5-1%); Oligocene and Miocene coccolith ooze and pelagic clay found in traverses I and II.

separated from the miopelagic clays by a thin belt of transitional siliceous-marly clay* with a CaCO_3 content of 10-50% and a $\text{SiO}_{2(\text{am})}$ content of 5-12%. The $\text{SiO}_{2(\text{am})}$ content in the coccolith-foraminifera oozes is from 3.28% to 5.25%, C_{org} is from 0.2% to 0.65%, Fe is up to 5.96% and Mn to 2.72% (after recalculation as carbonate- and silica-free, Fe is 13.5% and Mn 6.5%), and P is 0.05-0.12%.

To the east of the EPR are hemipelagic (mottled or greenish gray) oozes. In the Gulf of California area these are terrigenous sediments with a CaCO_3 content of 1-10%, $\text{SiO}_{2(\text{am})}$ is 1.0-3.7%, C_{org} is 1.0-2.11%, Fe is 3.68-9.66%, and Mn is 0.031-4.8%. In the Guatemala Basin the hemipelagic oozes are sometimes slightly siliceous ($\text{SiO}_{2(\text{am})}$ is 4.9-7.4%, and using the factor 1.6, 7.8-11.8%). They contain 0.0-30% CaCO_3 , 0.96-1.92% C_{org} , 0.03-0.04% P, 3.18-5.98% Fe, and 0.26-4.2% Mn.

In the center of the Guatemala Basin, on the east flank of the EPR, there are slightly siliceous, dark-brown miopelagic clays. They contain 0.0-13.7% CaCO_3 , 4.40-5.60% $\text{SiO}_{2(\text{am})}$ (with the factor 1.6, 7.7-9.0%), 0.28-0.96 C_{org} , 0.06-0.30% P, 4.18-7.48% Fe, and 2.28-6.66% Mn. These clays thus differ from the slightly siliceous miopelagic clays (radiolarian clay oozes) of the open Pacific (to the west of the EPR) by a markedly higher content of C_{org} and especially manganese (Tables 1-3). These may be the most manganiferous miopelagic clays in the entire Pacific Ocean. Hemipelagic, mottled and greenish-gray, slightly siliceous, terrigenous oozes of the Guatemala Basin grade into oxidized (Eh +400-600 mV, according to A. G. Rozanov), dark brown oozes with a large amount of coarse (up to 10-12 cm), rounded ferromanganese concretions and manganese microconcretions. Holocene and sometimes Pleistocene sediments were not present at many of the localities along traverses I and II. The sediment surface in those places consisted of oozes as old as early Oligocene (Fig. 3). Pre-Quaternary sediments are either zeolitic eupelagic clays or slightly siliceous to siliceous (radiolarian-clayey) oozes ($\text{SiO}_{2(\text{am})}$ 5-15%). The muds are highly enriched in Fe, Mn, and sometimes minor elements (Cu and Ni). In the Guatemala Basin the core contained only slightly

*Not shown on Fig. 2 because of the small scale.

TABLE 1. Content of Certain Elements and Components in Quaternary Eupelagic Clays, Traverse I (Pacific Ocean), %

Horizon, cm	CaCO ₃	C _{org}	Fe	Mn	P	SiO ₂ (am)
Sta. DM-3830-15 (depth 4,975 m, 10° 01' 4 N, 140° 09' W)						
0-2.5	3.5	0.44	4.37	0.26	0.26	6.2
2.5-5	<3	0.41	4.35	0.19	0.22	6.7
5-7.2	<3	0.39	4.46	0.18	0.21	6.7
7.5-10	<3	0.44	4.35	0.15	0.20	6.2
10-12.5	<3	0.33	3.84	0.18	0.19	6.2
Sta. DM-3830-18 (depth 4,930 m, 10° 00' 0 N, 139° 55' 2 W)						
0-2	<2.0	0.27	4.43	0.33	0.39	3.4
1-6	1.25	0.32	4.80	0.42	—	—
2-5	<2.0	0.27	4.27	0.42	0.37	3.4
10-15	<2.0	0.16	4.53	0.18	0.33	4.4
15-18	<2.0	0.37	4.75	0.95	0.39	3.7

TABLE 2. Content of Certain Elements and Components in Oozes of the Northern Part of the EPR, %

Horizon, cm	CaCO ₃	C _{org}	Fe	Mn	P	SiO ₂ (am)
Light reddish-brown, manganiferous, coccolith-foraminifera ooze (sta. DM-3845, depth 2,850 m, 7°36'7 N, 102° 47'1 W)						
0-5	65.0	3.45	2.59	0.51	0.05	3.2
5-10	68.0	3.29	2.48	0.48	0.06	2.8
15-20	63.0	3.11	3.15	0.66	0.06	3.0
40-48	62.0	2.74	2.74	0.71	0.06	2.6
Reddish-brown, limy-clayey radiolarian pelagic ooze (sta. DM-3847, depth 3,350 m, 7°25'2 N, 100°34'0 W)						
0-1	22.2	0.65	—	—	0.12	6.2
1-2	23.0	0.69	4.22	1.34	0.11	6.0
2-5	28.9	0.66	—	—	0.11	5.8
5-10	30.8	0.65	—	—	0.11	5.5
10-12	34.7	0.75	—	—	0.10	5.3
12-15	30.8	0.35	—	—	0.10	5.3

siliceous (diatom-radiolarian) Quaternary oozes. The most common Mn content on traverse G was from 0.5% to 5% (see sta. DM-41-3850 and -3851 on Fig. 3 and in Table 4). Mn extrema are accompanied extrema of Cu, Ni, and Zn.

Calcium Carbonate and the Calcite Compensation Depth (CCD). The CaCO₃ content in the upper layer (0-3 cm) of sediment is from 0.0 to 85.6%. If we ignore older sediments (Miocene and Oligocene), however, the ranges is reduced to 0.0-71.3% (Fig. 4). The carbonates are all CaCO₃; they have been shown by diffractometry to be mainly low-magnesian calcite. The Ca content (along with Mg, determined by atomic absorption techniques) is found to be closely related to CO₂. The Mg content in foraminifera oozes generally ranges from 0.28% to 1.00%, and in miopelagic clays and siliceous pelagic oozes, from 1.08% to 2.64%.

The distribution of CaCO₃ in Holocene and late Pleistocene bottom sediments is highly dependent on the present position of the CCD and bottom relief: carbonates are abundant where the CCD is below the sea floor, and sparse (<10%) where it is above. The lowest level of the CCD depth in the Pacific is known to be in the central part of the equatorial zone [9, 19, 20]. It becomes shallower toward the east [9, 16, 24]: in the western part of the Clarion-Clipperton area it is 4800 to 4900 m, in the area of traverses A and B of the DOMES expedition (see Fig. 1) it is 4400 m, and in the eastern part of the Clari-on-Clipperton area it is 3500-3600 m.

In order to determine the CCD more precisely, we have compiled a graph of the relationship between CaCO₃ and depth in the study area (Fig. 5). In the region of the EPR, the Holocene level is found to be at a depth of 4400 m. Where there is a large area of carbonate sediments (CaCO₃ > 50%) on the west flank of the EPR and on the Equator (140°E), the water depth is less than the CCD (for CaCO₃). A narrow belt containing 10-50% has been deposited on a topographic high caused by the Clipperton fracture zone [14].

TABLE 3. Content of Certain Elements and Components in Late Quaternary (mainly Holocene) Oozes of the Guatemala Basin, %

Horizon, cm	CaCO ₃	SiO ₂ (am)	C _{org}	Fe	Mn	P
Brown, manganiferous, clayey radiolarian miopelagic ooze (sta. DM-3849, depth 3,690 m, 6°48'3 N, 94°16'53 W)						
0-1	<2,0	-	0,54	-	-	0,18
0-2	<2,0	-	-	6,11	2,06	-
2-5	8,1	-	0,58	-	-	0,45
5-10	10,1	-	0,51	6,03	1,38	0,14
10-15	7,0	-	0,58	-	-	0,13
15-19	6,0	-	0,45	5,85	1,20	0,12
Brown, manganiferous, clayey radiolarian miopelagic ooze (sta. DM-3852, depth 3,580 m, 6°32'8 N, 92°47'3 W)						
0-2	<1,0	6,80	0,52	4,80	2,33	0,10
2-5	<1,0	5,29	0,56	4,69	4,96	0,11
5-8	<1,0	4,85	0,49	4,61	2,58	0,10
8-12	3,0	5,00	0,46	4,59	8,70	0,10
12-15	11,1	4,75	0,47	4,35	4,80	0,10
15-18	14,11	-	0,71	4,24	4,66	0,11
18-21	15,10	-	0,81	4,29	3,75	0,11
Brown, manganiferous, clayey radiolarian miopelagic ooze (sta. DM-3862, depth 3,735 m, 5°45'8 N, 92°03'13 W)						
0-3	9,6	6,20	0,28	4,76	2,29	0,30
3-6	13,5	4,90	0,28	4,64	7,20	0,18
6-10	18,3	5,30	0,67	4,80	2,34	0,18
10-15	21,2	-	0,62	3,77	2,20	0,12
15-19	25,1	-	0,58	3,64	1,90	0,13
Gray, clayey radiolarian hemipelagic ooze (sta. DM-3,900, depth 4,125 m, 12°16'8 N, 96°06' W)						
0-1	<1,0	5,00	1,63	4,02	2,40	0,29
1-5	<1,0	-	1,31	-	-	0,05
5-9	<1,0	-	1,54	-	-	0,04
9-14	<1,0	-	1,59	-	-	0,04
14-18	<1,0	-	1,46	-	-	0,04
18-20	<1,0	-	1,47	-	-	0,04

The position of the CCD on the east flank of the EPR and in the Guatemala Basin was found to be at 3400 m, some 1000 m higher than on the west flank of the EPR. Bottom samplers picked up Oligocene, Miocene, and Pleistocene sediments on traverses I and II. They were mostly eupelagic clays (with zeolites) and miopelagic clays (with radiolarians and diatoms). But pelagic nannoplanktonic (mainly discoaster) oozes were found at nine stations at the surface (rarely to depths of 1-2 m) of the sea floor. The CaCO₃ content in these oozes ranges from 80 to 90%, and C_{org} from 0.1 to 0.2%. These pre-Quaternary nanno oozes range in depth from 4700 to 4950 m. Oligocene and Miocene pelagic clays containing less than 1-3% CaCO₃ occur at these same depths. This indicates that the CCD varied by at least 500 m in the region of traverses I and II during the period from 37 up to 5 MY ago.

Oligocene and Miocene nanno oozes were originally deposited not at the depths where they are now found, however, but about 500 m shallower. This conclusion follows from the theory of plate tectonics, according to which the area of traverse II, at the time of deposition of the oozes, was located somewhere on the flank of the ancient EPR at a depth of 4100-4400 m. During the time from 30-35 MY ago until now, this area "left" the EPR and moved northward a considerable distance, to where it is now found. As it moved away, it subsided, in accordance with the Sclater curve. The difference between the original depth (when Oligocene oozes were deposited) and the present depth (where it was found on traverse II) is, according to our calculations, about 500-600 m. According to data in the literature [10, 27], the CCD in the equatorial Pacific (north of 4°N) 30-37 MY ago was at 4200-4300 m, and between 3°N and 3°S it was at 4800-5000 m [10], i.e. at about the same depths as at present.

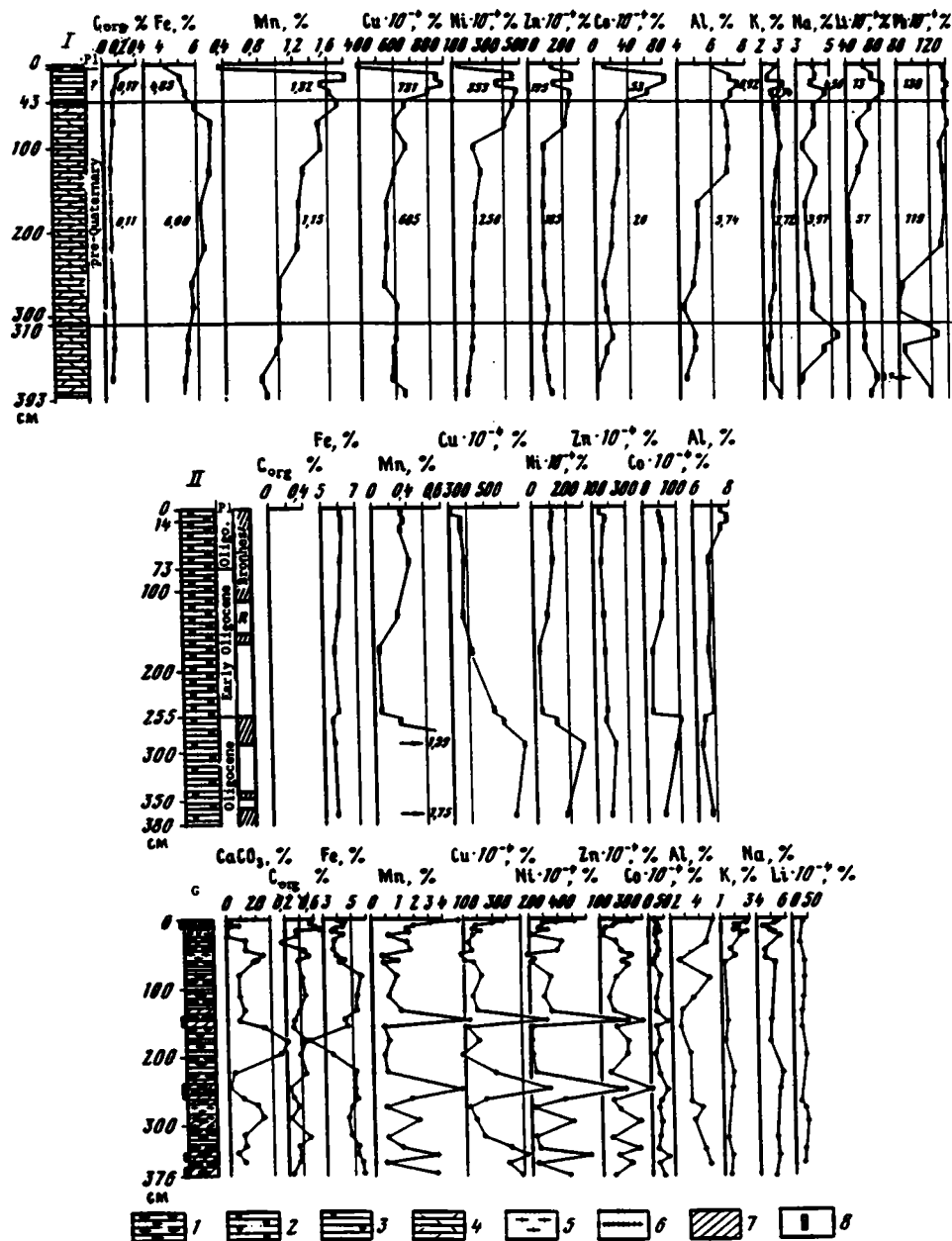


Fig. 3. Distribution of elements and components in the bottom sediments on traverse I (sta. DM-41-3830-28, depth 4940 m), traverse II (sta. DM-41-3833-5, depth 4910 m), and traverse G (sta. DM-41-3850-51, depth 3660 m). Ages from micropaleontologic data (V. V. Mukhina and S. B. Kruglikova), paleomagnetic ages determined by Yu. Yu. Ivanov (N and R are normal and reverse polarity, respectively). Age of sediments on traverse G is Quaternary. Sediment type: 1) reddish-brown, slightly siliceous miopelagic clays ($\text{SiO}_{2(\text{am})}$ 5-10% or 10-15% biogenic siliceous fossils); 2) same, but with less (~5%) siliceous fossils; 3) same, but with more ($\text{SiO}_{2(\text{am})}$ ~3%) poorly preserved siliceous fossils; 4) brown eupelagic clays, with zeolites; 5) brown miopelagic, slightly limy clays (CaCO_3 10-20%, coccoliths, foraminifera); 6) same, but with a large amount of coccoliths and foraminifera (CaCO_3 30-50%); 7) slightly manganiferous sediments (Mn 1-3%); 8) manganiferous sediments (Mn 3-6%).

Early Miocene oozes (about 18-19 MY in age) found on traverse II were deposited at greater depths than Oligocene oozes on the same traverse, since in the time span of 10 MY (the difference in ages of the sediments) the area moved a considerable distance away from the EPR and into somewhat deeper waters. At the time of deposition these depths were not greater than the CCD, however.

TABLE 4. Content of Certain Elements and Components in Quaternary Pelagic Oozes, Traverse G, % (sta. DM-3883, depth 3610 m, 6°30'1 N, 93° 00'1 W, brown, miopelagic, radiolarian, manganiferous ooze)

Horizon, cm	CaCO ₃	C _{org}	Fe	Mn	Cu	Zn	Ni	Co	Li	Rb
5-15	12,1	0,25	3,85	4,60	423	456	360	55	42	35
18-23	13,8	0,75	4,24	2,22	316	379	249	59	40	38
30-35	19,8	0,30	3,30	1,64	310	316	245	44	40	48
44-49	<1,0	0,17	3,01	1,23	169	183	146	22	34	64
49-54	<1,0	0,11	3,36	3,68	410	346	369	33	42	64
54-58	<1,0	0,11	4,70	3,27	447	355	391	52	40	50
63-67	<1,0	0,18	5,35	3,03	441	316	408	55	42	40
84-87	<1,0	0,24	6,73	2,03	419	331	360	59	50	42
92-96	<1,0	0,26	6,73	1,73	374	336	373	52	60	44
108-112	5,4	0,41	5,61	3,35	569	413	627	81	48	40
130-135	14,8	0,78	4,50	0,65	267	357	249	37	44	35
315-320	19,8	0,18	3,39	1,76	243	357	369	48	42	40
325-330	27,6	0,70	3,39	0,38	163	360	215	55	48	35
395-400	11,8	0,38	4,90	-	377	360	339	66	50	37

Thus higher CCD's were found in the equatorial Pacific in both the present and geological past eastward from the central ocean toward the foot of the continental rise of North and Central America. At the present time the difference is 1600 m. For the equatorial zone of all the oceans of the world, the difference for modern sediments would be 2500-2600 m: the CCD is 5900-6000 m in the Cape Verde and North American Basins in the Atlantic [14], and 3400 m in the Guatemala Basin.

Considerable variation in CaCO₃ (carbonate cycles) was found in Quaternary sediments of the EPR (see Table 4, Fig. 3). The content of this component in Pleistocene oozes ranges from 1% to 50%. This confirms the view that the CCD in the equatorial Pacific is a function not only of the longitude (a decrease in depth from west to east) [10, 20, 21], but also cyclic changes in the climate and, apparently, changes in the properties of the seawater [10, 22, 23].

Amorphous Silica. In the upper layer (0-3 cm) of the sediments the SiO_{2(am)} content ranges from 1.3% to 7.7%, which using the factor 1.6 (approximately 1.6 times reduction in the results because of incomplete dissociation of the siliceous fossils, mostly thin-shelled radiolarians) becomes 2.1% to 12.3%. As previously known [12, 13], SiO_{2(am)} has a zonal distribution. Low-silica clays (SiO_{2(am)} < 3%), which are found to the north of the region, become more siliceous at about latitude 15°N. The average content on traverse C was 1.8%, and on A and B, 10.8% and 11% respectively (these data were not obtained by direct determination of SiO_{2(am)}, however, but by calculations based on the amount of siliceous fossils in the sediments). Recall that silica deposits in the eastern equatorial Pacific were found to have about the same amount of SiO_{2(am)} (5-10%) [13]. Thus, sediments in the study area, calculated on a carbonate-free basis, generally contain 5-10% amorphous silica. These are consequently not siliceous pelagic oozes. It is therefore incorrect to call these radiolarian oozes or clays. The statement by Sval'nov [18] that amorphous silica averages 60% in radiolarian sediments, 28% in radiolarian clays, 32% in radiolarian, limy oozes (CaCO₃ up to 30%), and 31% in coccolithradiolarian oozes (CaCO₃ up to 50%) does not agree with our data or that of other authors [12, 13, 15, 26]. Sval'nov's data was obtained by calculation of silica particles in smears, but these data generally give results that are far too high.

Only those sediments on traverses A, B, and I that contain 10-15% SiO_{2(am)} (using the factor 1.6) can be called slightly siliceous (see Fig. 4). Typically siliceous (radiolarian) sediments are located only in the equatorial zone (150°E), where SiO_{2(am)} is as much as 25.2-37.6% [12].

The siliceous fossils in Quaternary oozes on traverses I and II are mostly diatoms (mainly *Pseudoeunotia doliolus*). There are lesser amounts of modern radiolarians (*Gollosphaera tuberosa*), as well as redeposited Miocene and Pliocene radiolarians. In separations from aqueous suspensions in the study area (5-7-m layer), the most common diatoms encountered include *Thalassionema nitzschioides* (dominant), *Th. sp.*, *Coscinodiscus nodulifer*, *Nitzschia bicapitata*, *Rhisosolenia styliformis*, and *Pseudoeunotia doliolus* (identifications by V. V. Mukhina). Altogether 25 species of diatoms, belonging to 16 genera, and two species of silicoflagellates, were found in the study area.

The amount of siliceous fossils decreases somewhat with an increase in depth in the sediment. This tendency in Pacific sediments has been noted previously [26].

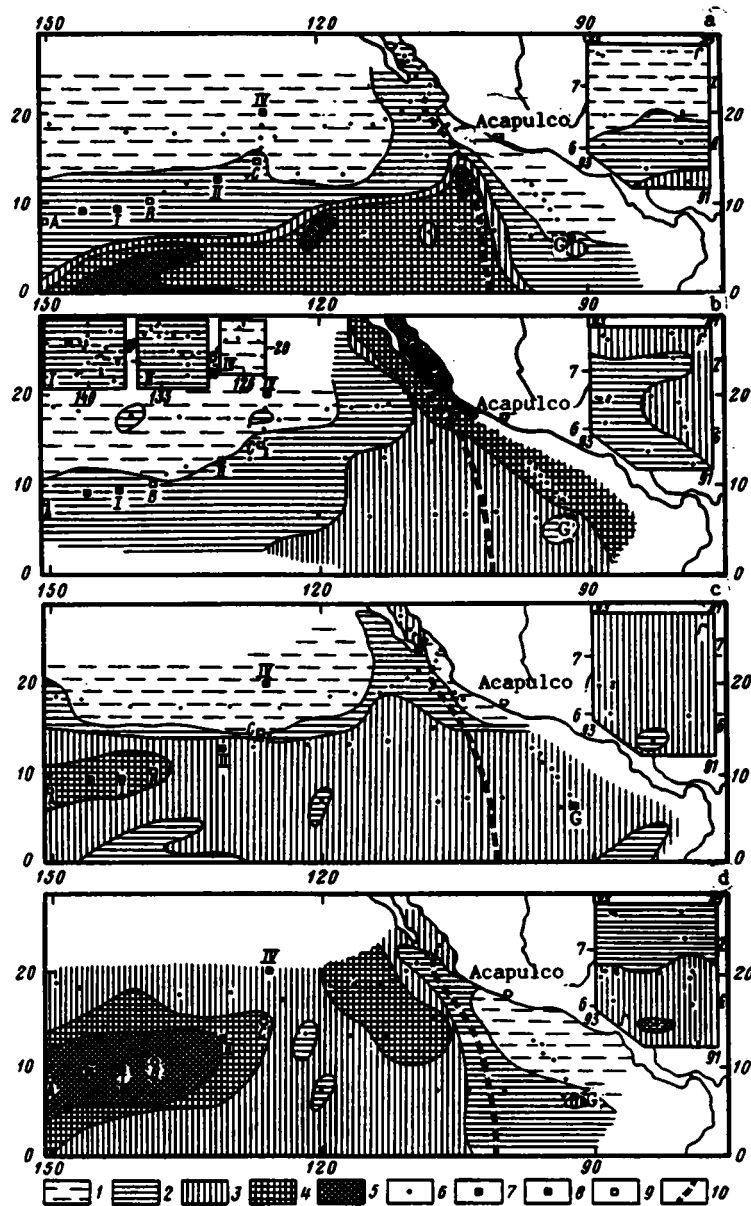


Fig. 4. Maps showing the distribution of CaCO_3 (a), C_{org} (b), $\text{SiO}_{2(\text{am})}$ (c), and P (d) in the upper layer (0-3 cm) of bottom sediments. CaCO_3 content, %: 1) <1; 2) 1-10; 3) 10-50; 4) 50-70; 5) 70-85. C_{org} content, %: 1) <0.25; 2) 0.25-0.50; 3) 0.50-1.0; 4) 1-2; 5) >2. $\text{SiO}_{2(\text{am})}$ content, %: 1) <2; 2) 2-5; 3) 5-10; 4) >10. P content, %: 1) <0.05; 2) 0.05-0.1; 3) 0.1-0.15; 4) 0.15-0.20; 5) 0.2-0.3. 6) Station; 7) traverses of expeditions DM-41, DOMES, and DM-2483, respectively; 8) crest of EPR; G, traverse in Guatemala Basin; I, II, IV, and G, traverses of expedition DM-41.

This conclusion is basically confirmed by our calculations from smears as well as layer-by-layer determination of $\text{SiO}_{2(\text{am})}$ in bottom samples (0-20 cm layer) (see Tables 2 and 3). Early Miocene miopelagic clays (sta. DM-3830-4, depth of 4890 m) yielded 0.29-0.43% $\text{SiO}_{2(\text{am})}$, i.e. 10-15 times less than in Pleistocene oozes in the same area. $\text{SiO}_{2(\text{am})}$ was not determined chemically in Oligocene sediments. Judging by smears studied, the silica content (radiolarians and diatoms) in Oligocene pelagic clays and coccolith oozes usually does not exceed 5-10% of the sediment.

Organic Carbon. The upper layer of sediment in the study area was found to contain 0.12-1.92% C_{org} . Only in a sample of foraminifera ooze from the crest of the EPR (sta. DM-3845, depth of 2850 m) was as much as 3.45%. A content of >2% (up to 7.8% in one sample) is limited to the Gulf of California.

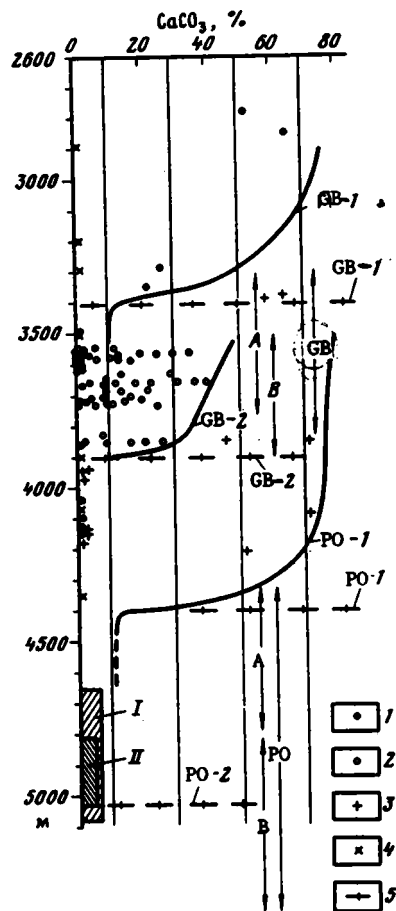


Fig. 5. Calcite compensation depth (CCD) in the northern part of the eastern equatorial Pacific. Samples: 1) Guatemala Basin (layer 0-3 cm, mainly late Holocene); 2) same, layer 25-400 cm (Pleistocene); 3) open Pacific (west of the EPR); 4) samples from expedition DM-9 (see Fig. 2), layer 0-3 cm. 5) CCD: GB, Guatemala Basin (GB-1, Holocene; GB-2, individual stages of the Pleistocene); PO, Pacific Ocean west of the EPR (PO-1, Holocene; PO-2, Oligocene); I and II, samples from traverses I and II. Vertical arrows indicate depths most favorable for development of ferromanganese nodules: GB, in the Guatemala Basin (A, Holocene; B, Pleistocene); PO, in the open Pacific west of the EPR (A, Holocene; B, Oligocene). Sample areas outlined by curved lines.

Examination of the maps (see Fig. 4) and the data in Table 5 suggests two definite conclusions. Firstly, the content of C_{org} decreases from the shelf to the open ocean, most markedly in a belt at 15-20°, i.e. beyond the limits of the highly productive equatorial zone. Southward areas of C_{org} content of 0.25-0.5% and 0.5-1% extend along the equator. Secondly, the C_{org} content in the open ocean (beyond the continental slope of California and west of the EPR) increases from the north (20°N) to the south (10°N).

The data in Tables 1-5 indicate that in neither the upper layer (0-20 cm) nor in the entire Quaternary section is there a clear-cut downward decrease in C_{org} . Thus active decay of organic matter ceases after burial (by 1-2 cm of sediment). The generally minor changes in C_{org} content in the column are apparently caused by changes in the climate and in the properties of the water (and possibly by intermittent exposure of the organic matter at the surface of the sediment, or by rapid sedimentation).

TABLE 5. C_{org} Content in Bottom Sediments, %

Station, transverse	Depth, m	Horizon, cm	Age	C _{org} , %	
				range	average
I. Upper layer of sediment					
IV	4555-4750	0-3	Q _{IV}	0,11-0,17	0,14
II	4810-5030	0-3	Q _{IV} -III	0,30-0,40	0,37
I	4650-5080	0-3	Q _{IV} -III	0,23-0,46	0,33
G		0-3	Q _{IV}	0,22-0,96	0,50
II. Pleistocene oozes (clays)					
a. Guatemala Basin					
DM-3899	4095	42-47	Plt	3,00	-
DM-3897	3960	10-400	"	0,30-2,89	-
DM-3892	3825	42-47	"	2,68	-
DM-3863 (tra- verse G)	3700	10-400	"	0,19-0,48	-
G (entire traverse)	3490-3861	10-500	"	0,15-1,0	0,30
b. EPR					
DM-3845	2850	0-120	"	2,63-3,68	-
Entire sec- tion of EPR	2850-4090	10-400	"	-	0,30
c. Area between traverses I and II					
DM-3832	4887	0-25	HI-Plt	0,10-0,27	-
		25-510	Plt	0,10-0,17	-
DM-3831	5015	0-30	HI-Plt	0,19-0,29	-
		40-345	Plt	0,15-0,19	-
III. Pre-Quaternary (Miocene and Oligocene) nanno oozes and pelagic clays					
DM-3830-28	4940	10-43	N	0,10-0,17	0,13
(traverse I)		44-375	"	0,08-0,12	0,10
DM-3840-43	5040	43-300	"	0,04-0,10	0,08
(traverse I)					
DM-3833-14	5046	20-400	P ₁	0,10-0,15	0,12

Such changes are of course typical of the Quaternary period. In the study area there were no significant changes in the processes of organic matter supply and deposition in the Pleistocene compared to the late Holocene.

In pre-Quaternary nanno oozes and pelagic clays the C_{org} content is significantly less than in Quaternary sediments in the same area (see Fig. 3 and Table 5). This suggests two hypotheses: 1) the bioproductivity in the upper layers of the water was less during the Miocene and Oligocene than at present; and 2) 18-35 MY (the age of the sediments) was sufficient time for decay of much of the buried organic matter during diagenesis. It is possible that both of these circumstances contributed to the lower C_{org} content in pre-Quaternary sediments.

Phosphorous. In 18 separation suspension samples (5-7 m layer) from the study area we found 0.12-0.420% (average 0.25%) phosphorous, while in the upper layer of bottom sediment it was 0.03-0.42%. The lowest content (0.03-0.04%) was characteristic of hemipelagic oozes of the Guatemala Basin, and the highest (0.24-0.42%) was found in oozes of traverse I (see Fig. 4), i.e. the area with the highest sedimentation rate. The highest content of phosphorous in the oozes is associated with large amounts of fish bones and teeth (mostly sharks). Near the continent, where sedimentation is rapid, bones and teeth are sparse and generally not found. As in the pelagic regions of the Atlantic [6], the phosphorous here is apparently associated mainly with finely dispersed minerals. The phosphorous content in the section reaches as much as 0.48% (Table 6), with the higher content (>0.20%) found both in Paleogene and Neogene and in late Quaternary clays. Large amounts of phosphorous are mainly typical of beds that were deposited slowly. The notable decrease in phosphorous content downward in the late Quaternary clays in the Guatemala basin (see Table 3) can be explained by partial solution of phosphatic fossils with burial.

* * *

TABLE 6. Distribution of Phosphorous in the Section, %

Station	Depth, m	Horizon, cm	Type of sediment	Age	P
Traverse I					
DM-3830-28	4940	0-4	Siliceous pelagic ooze	Plt	0,42
		10-43	Miopelagic clay	N	0,43-0,46
		44-375	Eupelagic clay	"	0,29-0,48
DM-3830-43	5040	0-43	Siliceous pelagic ooze	Plt	0,29-0,36
		44-295	Miopelagic clay	N ₁	0,24-0,36
DM-3832	4887	0-25	Siliceous pelagic ooze	Plt	0,07-0,11
Traverse II					
DM-3833-5	4910	0-14	Same	Q	0,09
		15-255	Miopelagic clay	P ₁	0,09-0,18
		256-380	Eupelagic clay	"	0,20-0,23
DM-3833-14	5046	0-25	Siliceous pelagic ooze	Q	0,09-0,10
		25-400	Miopelagic clay	P ₁	0,10-0,45
DM-3833-22	4760	0-13	Siliceous pelagic ooze	Plt	0,21
		14-71	Miopelagic clay	N ₁	0,32-0,42
		71-153	Coccolith and radiolarian ooze	"	0,08-0,24
Section across the EPR					
DM-3836	4210	0-372	Coccolith-formaminifera limy clayey ooze		0,06-0,19
DM-3839	3850	0-408	Same	"	0,08-0,16
DM-3846	3280	0-270	Limy-Clayey ooze	"	0,08-0,10
Guatemala Basin					
DM-3850	3660	10-376	Miopelagic siliceous, locally limy clay	"	0,04-0,12

High biological productivity of the water on the west coast of North and Central America results in a significant C_{org} content in the oozes of the hemipelagic areas of the ocean, while the high biological productivity of the water in the equatorial zone results in a high content of organic matter in the equatorial region.

Separation suspensions we collected during expedition DM-41, as well as previously published information from the study area near the equator, show an increase in suspended sediment and relatively large amounts of C_{org} (11.0-21.1%), $CaCO_3$ (10-30%), and $SiO_{2(am)}$ (5-15%). This sediment is consumed in the upper layers of the sea by zooplankton and nekton and returned to the water as pellets, which are rapidly, and thus with little change in form, deposited on the sea floor [6, 13]. This is also confirmed by the sediment traps we used on traverses II and G. It is as if the high content of $CaCO_3$, $SiO_{2(am)}$, and C_{org} in suspended sediments of the upper layers of the water are projected onto the sea floor. As a result, in the equatorial area (and

for organic matter, the continental margins as well), the zones of high content of these biogenic components in suspended sediments and in bottom sediments virtually coincide.

In the northern part of the Guatemala Basin, the C_{org} content near the base of the continental slope is 2-4 times greater than at the same depth (3500-3700 m) away from it. This is caused mainly by the effects of near-shore upwelling, which is lacking in the open ocean; and also by the rapid rate of sedimentation and an environment conducive to the preservation of organic matter on the sea floor at the continental slope. A decrease in the C_{org} content toward the center of the Guatemala Basin is the result of a gradual facies change from greenish-gray hemipelagic ooze to brown, oxidized, manganiferous miopelagic ooze (clay). These facies may contain the similar amounts of $CaCO_3$ and $SiO_{2(am)}$.

To a large extent, the distribution of $SiO_{2(am)}$ depends on climate zones: it increases from north to south as far as the equator. The highest gradient in the increase in $SiO_{2(am)}$ content is between 10° and $15^\circ N$. In the equatorial zone to the south, there is an increase in amorphous silica. This is related to an increase in biological productivity in the zone of equatorial divergence [13].

Unlike $SiO_{2(am)}$, calcium carbonate conforms to a strict law of vertical zonation, i.e. the position of the CCD.

Most of the bottom waters of the Pacific are formed in the Antarctic region, moving northward as slow bottom currents to the equator, where they turn to the northeast and east until they enter the northeastern equatorial region [10, 13]. Inasmuch as they are completely isolated from surface waters and are not intermixed, they arrive in our study "old" and corrosive [1]. Because of this circumstance, the deep waters in both the northern part of our study area and in the entire northern Pacific are markedly undersaturated with respect to calcium carbonate. As a result, the bottom sediments in the pelagic regions of the ocean are almost completely devoid of $CaCO_3$ below the CCD [12, 13].

Deep oceanic waters that are saturated with calcite to about 60-70% are evidently found at the base of the continental slope off California and on the west flank of the EPR. The higher level of the CCD in these eastern Pacific regions compared to the open ocean is due to the active supply of organic matter to the sea floor and its presence at the sediment surface there. The additional amounts of CO_2 formed during this process becomes available in the bottom waters, increasing the pCO_2 and thus shifting the CCD upward.

The "old" and corrosive (Antarctic) bottom waters evidently do not enter the Guatemala basin for two reasons: 1) because they are cut off by the EPR, and 2) because of the relative shallowness of the basin (it is 600-700 m shallower than the CCD level on the east flank of the EPR). Consequently, the reason for the corrosiveness of the bottom waters of the Guatemala Basin should be different from that of the open ocean west of the EPR.

The reason for the very great depth of the CCD in the central equatorial Pacific (>5000 m) [20, 21]) is the high biological productivity of the waters and the active supply of biogenic (including carbonate) material to the sea floor. In the Guatemala basin, however, the average annual production and supply of biogenic material to the sea floor is several times greater than in the open ocean, while the CCD is 1600 m higher than in the center of the ocean. Thus, far offshore there is a direct relationship between the level of the CCD and biogenic productivity, while near the western coast of North America, there is not. Whence this paradox? In the coastal regions of the Pacific, unlike the central ocean, there is a vigorous supply of organic matter to the sea floor, as well as its decay there, associated with the active accumulation of corrosive (with respect to $CaCO_3$) CO_2 in the waters. Recall that in the study area there is a strong vertical gradient in pH and pCO_2 in the upper part of the intermediate layer of the water: $pH = 0.2-0.4$ and $pCO_2 = 2.5 \cdot 10^{-4}$ atm [8]. This layer (200-1000 m) has minimal values of pH (<7.7) and O_2 (0.10-0.05 ml/liter) and maximal pCO_2 $((10-12) \cdot 10^{-4}$ atm) and ALK/CL (>0.128).

Thus, the corrosive effect of the water on carbonate begins in the layer of acid minimum (i.e., on the continental slope). This appears especially clearly under the eastern continental current [23].

Below the oxygen minimum, i.e. at depths of 2000-3000 m, the pCO_2 decreases by an average of 5.6 to $6.4 \cdot 10^{-4}$ atm, and the pH increases to 7.90 [1, p. 115]. Consequently, the corrosiveness of the water is reduced and the calcite saturation increases to as much as 90%. Below 3000 m the corrosiveness of the water increases again. This increase is the result of higher pCO_2 and lower pH due to CO_2 being supplied to bottom waters by the decay of organic matter on the sea floor. Apparently, pore water is also corrosive with respect to carbonate. In our view, this is due to the high level of the CCD in the eastern equatorial Pacific. As for the Guatemala Basin, we suggest that there are two other features that contribute to the corrosiveness of the bottom waters: 1) poor exchange (ventilation) of deep waters because of isolation of the basin from the open ocean by volcanic ridges and uplifts, and 2) the introduction of large amounts of hydrothermal materials here (solutions and gases). The latter is indicated not only by direct observation of hydrothermal vents [13, 15, 25], but also by the enrichment of sediments in the Guatemala Basin, EPR, and Gulf of California in manganese, iron, lead, zinc, and other microelements.

The solution of calcareous material in corrosive waters requires months and years. Planktonic foraminifera tests, which we submerged in Karponov nets to automatic stations on the sea floor along traverses I and G (below the CCD), after 36 and 11 days respectively were only reduced to crystallites, but were not dissolved. Consequently, most foraminifera tests and nannoflora are not dissolved in the water column, but on the sea floor, under the effects of corrosive bottom waters and pore water, as suggested in the literature [21, 24].

Changes in the level of the CCD during the Cenozoic evolution of the ocean has long been established [10, 20, 24]. The principal causes of these changes have been changes in the climate and circulation of the water. Carbonate cycles lasted 10,000-40,000 years in the Pleistocene [5, 24] and about 500,000 years in the Paleogene and Neogene. This cyclicity is apparently also explained by the deposition of both coccolith oozes and eupelagic and miopelagic clays during short intervals of time in the Oligocene and Miocene in roughly the same areas of the sea floor (traverses I and II).

The distribution of phosphorous differs somewhat from that of the three biogenic components mentioned above. The highest concentrations in the open ocean are limited to areas where sedimentation is very slow or completely lacking. The greatest amount of biogenic remains are found here, in the uppermost layer of sediment. The high P content in the region of 10-20°N, 110-120°E is apparently due to the accumulation of such remains.

Of particular interest is the possible effect of hydrothermal activity along the EPR, or even the Clarion and Clipperton transform faults, on biogenic pelagic sedimentation, as mentioned from time to time by various authors. The information presented herein shows that all of the distribution patterns of the biogenic components in sediments can be explained by exogenic factors, which can be summed up in a concept of a broad (climatic), circumcontinental and vertical zonation of sedimentation. Introduction of additional biogenic elements (mainly SiO₂) by hydrothermal sources had no appreciable effect on bioproductivity. This is understandable, since the hydrothermal sources discharge into bottom waters at depths of 2-3 km or more, and the introduction of any components from the earth's interior into the zone of photosynthesis requires a system of vertical circulation (divergence, upwelling), which essentially does not exist in the region of the Pacific under study.

The possible effects of changes (produced by hydrothermal activity) in the physical and chemical parameters (pH, pCO₂) of bottom waters on the solubility of carbonates, i.e. the position of the CCD, is not entirely clear. Although as shown above the sharp rise in the CCD to the east of the EPR is due to the increase in organic matter resulting from high bioproductivity, in the western part of the Guatemala Basin some effects from hydrothermal activity are possible as well. This is supported by the high bathymetric level of the CCD in the entire eastern Pacific, where bioproductivity is low (such as in the Bauer Basin). The latter may be explained otherwise, however: the introduction of "old", strongly transforming, CO₂-enriched Antarctic bottom waters.

The suspensions in study area contained a remarkable quantity of various metals (including Fe, Mn, Cu, Ni, and Co). They fall to the sea floor with the biogenic material and serve as one of the sources of ore in bottom sediments for the formation of ferromanganese nodules. This is one of the reasons for the high concentration of such nodules throughout the highly productive equatorial Pacific [8, 26]. In areas where hydrothermal sources supplement biogenic material in providing ore components, the nodules "grow" more rapidly: concretions lying on Pleistocene oozes in the basin reach 10-12 cm in diameter.

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Characteristics of products, resulting from three Precambrian hypergenetic processes — surface (weathering horizons, caliche), underwater (halmyrolysis products), and subsurface (paleo-ground water horizons, zones of acid dissolution) hypergenesis — have been established. Three Precambrian evolution stages were recognized: Early and Late Archean, Late Proterozoic.

The concept of "hypergenesis" expresses the combination of chemical, physical and biochemical processes and phenomena that take place in the near-surface portion of the Earth's crust, the zone of hypergenesis. This is the zone where the lithosphere interacts with the atmosphere, the hydrosphere and outer space. The 1985 characterization of hypergenetic products [17] listed three hypergenetic process categories and corresponding products: surface, underwater (halmyrolytic) and subsurface (the "deep" category), according to Kravchenko [12]. The products of each of these hypergenetic types are characterized by specific locations and distribution areas [16, 17].

The survey of lithogenetic evolution problems throughout Earth history [16] led us to a clearcut conclusion on the inevitability of essential changes through time in all hypergenetic categories and their products.

These changes were especially intensive on continental surfaces where weathering horizons of different composition developed, along with products of their reworking.

Significant disagreements exist in the literature, however, on the evolution of hypergenesis and its component, the weathering horizon development [2, 9, 10, 21, 24, 29]. These different views are briefly reviewed in the following. A large group of workers, headed by Academician A. V. Sidorenko and including Yu. A. Borshchevskii, O. S. Koryakin, E. A. Kulish, O. I. Luneva, O. M. Rozen, Sv. A. Sidorenko, V. A. Tenyakov, V. M. Chaiki, et al., developed ideas on the "essential" similarity between geochemical factors throughout known Earth's history [24, p. 10]. This led to the view of the essential similarity between hypergenic minerals, formed in the Precambrian and Phanerozoic, as well as a corresponding ability to use the same research criteria in the resource evaluation of beds of different ages.

An other group of authors believed that hypergenetic ore development conditions varied through Earth history. For this reason no hypergenetic deposits existed in the Precambrian and even the Lower Paleozoic similar to those found in younger systems (e.g., bauxite-bearing lateritic blankets, hypergenetic Ni, Co, Fe, and other ores [15, 17, 30, etc.]).

According to researchers of the first group, one basic indicator of the alleged "essential similarity" noted in the Precambrian was "the abundance both of redeposited and in situ eluvial deposits, as well as the total absence of even minor exotic hypergenic product" ([28], p. 57). The supposedly common development of weathering horizons in the Precambrian and "the established, essential similarity between Precambrian and Phanerozoic weathering horizons (based on the makeup of eluvial profiles and rock composition)" ([24], p. 9) were additional arguments to support the theory.

An alternate theory for the evolution of exogenic ore formation throughout Earth history [15, 17, etc.] claimed that intensive chemical differentiation of matter under conditions of surficial hypergenesis commenced only in the Devonian. During pre-Devonian times, especially during the Precambrian subsurface and partly underwater types of hypergenes were more common. These were associated with exogenic (subsurface and hydrothermal sedimentary [16, 17]) Pb, Zn, Cu, Au, U, and other ore deposits.

Because these assumptions have been disputed by many workers and because they are essential in the review of Precambrian hypergenesis problems in general, in the following we briefly summarize the comparative characteristics of post-Silurian weathering horizons and describe all three Precambrian hypergenetic categories.

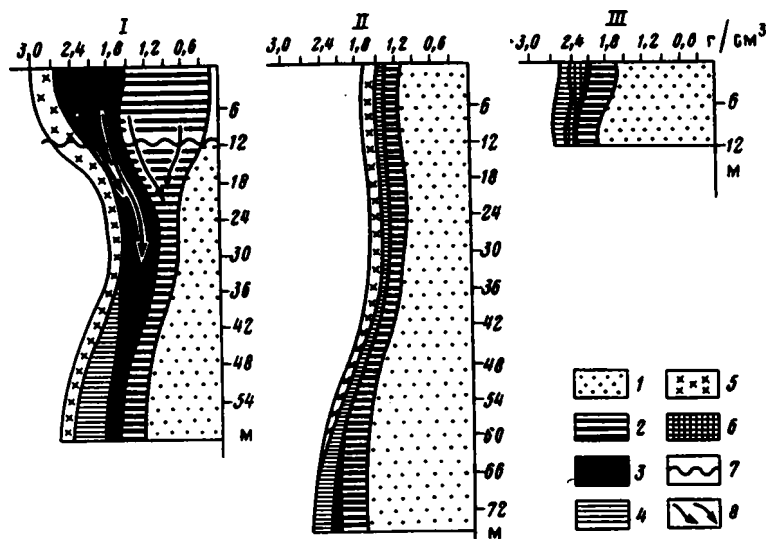


Fig. 1. Geochemical diagrams of weathering horizons. I) laterite profile, Republic of Guinea (P3-Q); II) kaolinitic profile, Ukrainian Shield (K2); III) "Precambrian weathering horizon." Acid leaching zone, Saltic Shield (PR1).

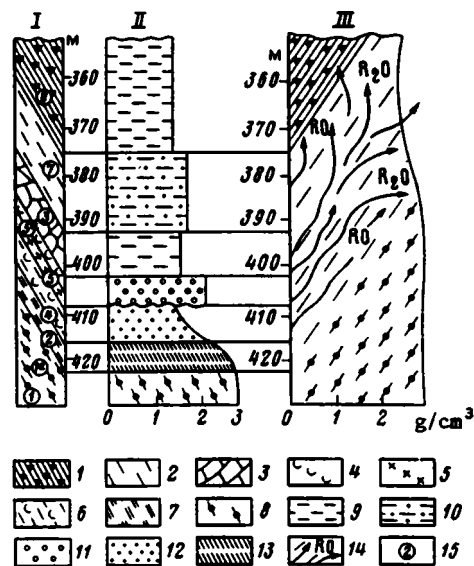


Fig. 2. Section interpretation, Drillhole #5257 (Priiskol'sk Mine KMA). I) lithological section; II) interpretation by supporters of a Precambrian age of the weathering horizon; III) interpretation by supporters of acid leaching theory (subsurface hypergenesis). 1) quartz-biotite schist with garnet; 2) quartz-sericite schist; 3) cataclastic quartz-chlorite rocks; 4) quartz with rare sericite flakes; 5) quartz-feldspar rock; 6) sericite quartzite; 7) mica-biotite mica schist, with quartz and chlorite; 8) amphibolite; 9) clay; 10) aleurolite; 11) quartz conglomerate; 12) kaolinite weathering zone; 13) illitic weathering zone; 14) acid leaching zone (subsurface hypergenesis); 15) sample number (Tables 2 and 3).

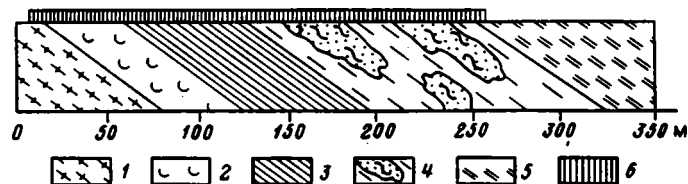


Fig. 3. Section of basal horizons of the Keivsk Series (Kola Peninsula, A2) acid leaching zone (subsurface hypergenesis). 1) biotitic plagioclase gneisses (A2); 2) quartzite with garnet; 3) quartz-kyanite and quartz-staurolite schists; 4) muscovite-quartz schists with plagioclase relicts; 5) biotite-muscovite schists; 6) acid leaching zones.

DISTINGUISHING CHARACTERISTICS OF POSTSILURIAN WEATHERING HORIZONS

Poorly stratified, very well developed thick weathering horizons are common in Postsilurian deposits of the USSR. The horizons formed during the Middle and Late Devonian, Early Carboniferous, Late Triassic, Early and Late Cretaceous (Albian—Turonian) intervals. Each period was characterized by a certain distribution area and partly by characteristic composition of the final weathering products. During Oligocene—Quaternary times, areas of chemically deeply weathered horizons developed in humid tropical zones beyond the USSR boundaries [17].

All these periods were distinguished by the development of kaolinitic and lateritic profiles, representing the final stage of rock decomposition on the surface under hot, humid climates. The profiles were tens, occasionally hundreds of meters thick (Fig. 1).

Rock density sharply declines both in the kaolinitic and the lateritic weathering profiles, occasionally to 50-75% of the original density. Silica, alkali- and alkali earth elements have been leached out intensively. Autometasomatic loss of Al, iron oxide and Ti usually occurs in the lateritic horizons, due to decomposition of surficial "kiras" features (Fig. 1). In the thick kaolinitic profiles the absolute Al content remains the same per volume unit, or decreases slightly (Fig. 1). Huge volumes of matter were transported in regular and colloidal solutions during these gradational processes. Differentiation of the solutions resulted in formation of large and usually rich hypergenetic deposits of various mineral resources.

It is important that the lower boundary of the Postsilurian weathering horizon usually is clearly marked by intensive clay development on the source rocks. More exactly: by rock hydrolysis, accompanied by the leaching out of alkali, alkali earth elements and reduction in rock density. Partial chloritization, saussuritization, pelitization, and sericitization of the source rocks were not observed, as they occur at depths of many hundreds of meters, even several kilometers. For example, in the record-deep Kola Peninsula drillhole, the Proterozoic and Archean rocks at certain depths (in particular, at 7680, 10,000 and 10,500 m) presently occur in cataclastic zones. These zones are several tens of meters thick, occasionally reaching a thickness of 100 m. "Hypergenetic processes did commonly occur in the rocks along fractures in contact zones (feldspar pelitization and amphibole saussuritization)" ([14], p. 51). Additionally, chlorite and sericite also occurred at such locations. These are essentially synchronous products of the same chemistry and are essential in formation of exogenic mineralizations in large parts of the Earth's crust.

Thus, one should commonly include tens, rarely hundreds of meters thick rock beds into weathering horizons during the analysis of post-Silurian sections. These, while of essentially different composition, did preserve the relict structures of the substrate. Hydrated Al-silicates (illite, montmorillonite, and kaolinite groups), oxides and hydroxides of Fe, Ti and Al, etc. hydrolyzite elements predominate in the mineral spectra of weathering products. The newly forming rocks were of significantly lesser density. These alterations generally were accompanied by hypergenetic-metasomatic processes that significantly distorted the general pattern of hypergenesis (Fig. 1).

GENETIC TYPES OF PRECAMBRIAN HYPERGENESIS

Efforts to establish the idea of "essential similarity between geochemical hypergenetic factors throughout the Earth history" ([24], p. 10) for many years had been based on the search for weathering horizons and their reworked products in the Precambrian deposits. The possibility that other hypergenetic formations may also occur among these ancient deposits has not

TABLE 1. Chemical Composition of Products of Halmyrolysis and "Ancient Weathering Horizons". Composition Percentages in Basic Rocks*

Sample number	Rock	Age	Sample location	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O ⁻	H ₂ O ⁺	P ₂ O ₅	Calcina- tion loss	Σ
1	tholeiitic	Q	Indian Ocean, 4800 m depth	50,56	0,71	14,08	0,61	9,63	0,17	10,60	11,76	1,50	0,08	0,04	0,16	0,06	0,38	100,34
2	Basalt halmyrolysis products (palagonitized basalt)	Q		40,12	0,89	17,90	11,60	0,28	0,20	2,50	0,44	1,80	2,72	12,56	3,20	0,08	5,20	99,49
3	Biotite-amphibolite schist	A	Krivoi Rog, "Severnaya" (North) Mine	51,49	0,77	16,27	1,12	7,02	0,27	4,45	10,67	0,40	3,26	-	-	0,06	5,71	101,49
4	Biotite-sericite schist	PR ₁		51,45	1,02	20,33	2,42	11,64	0,12	2,52	0,33	0,22	6,18	-	-	0,13	4,18	100,54
5	Mica-garnet schist	PR ₁		46,00	1,14	22,91	2,20	12,42		2,16	0,27	0,24	7,74	-	-	0,14	4,29	99,51
6	Amphibolite	A	Krivoi Rog	51,90	0,70	14,95	1,93	6,85	0,25	5,28	10,23	1,22	3,14	-	-	0,06	2,99	99,50
7	Quartz-amphibolite rock	PR ₁		58,60	1,37	17,00	2,03	6,15	0,05	4,86	0,65	0,34	6,35	-	-	0,08	2,65	100,13
8	Sericite-quartz-biotite rock	PR ₁		57,06	1,54	18,24	2,00	6,19	0,10	3,64	0,20	0,15	6,90	-	-	0,18	3,65	99,85

*Data on Samples 1 and 2: from publication [7] (p. 283); Samples 3-5: from publication [2] (p. 168), Samples 6-8: from [8] (p. 145).

TABLE 2. Chemical Composition of Rocks in the Weathering Horizon, %

Sample number	Rock	Density, g/cm ³	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	Calcin. loss	Σ
1	Amphibolite	-	50,85	0,65	15,86	1,45	10,21	0,15	9,22	4,44	4,04	1,02	Not det'd.	Not det'd.	2,34	100,23
1 ^a	Clear amphibolite	2,92	49,47	0,73	16,61	3,15	8,96	0,18	6,73	9,45	2,16	1,01	0,06	"	1,74	100,25
2	Biotite-sericite Schist with chlorite	-	48,36	0,82	17,32	3,56	8,24	0,15	7,60	7,33	2,87	2,95	0,08	"	1,73	101,01
4	Quartz-sericite schist	2,75	80,23	0,03	12,93	0,48	Not fd.	Not fd.	0,50	Not det'd.	0,39	3,80	Not fd.	"	1,78	100,05
5	Quartz-feldspar rocks	2,70	71,51	0,58	10,32	4,25	2,38	0,04	1,04	0,15	5,05	0,38	0,11	0,02	3,54	99,37
5 ^a			70,43	0,06	16,26	0,86	0,14	Not fd.	0,35	0,20	1,97	8,20	Not fd.	Not det'd.	1,20	99,67
7	Quartz-chlorite-sericite schist		55,15	0,32	10,90	6,47	12,80	0,22	5,51	0,43	0,53	5,28	"	"	1,94	97,61
8	Quartz-biotite schist	2,85	58,30	0,61	15,18	2,55	11,01	0,13	3,86	0,71	4,81	0,15	0,10	0,01	2,04	99,46

TABLE 3. Semiquantitative Spectral Analysis Data of Rocks*

Sample number	Rock	K	Na	Ba	Ti	Mn	V	Cr	Co	Ni	Zr	Nb	Se	Ce	La	I	Yb	Be	Li	Mo	Cu	Pb	Zn
8	Quartz-biotite	3	2	0,02	1	0,15	700	300	70	300	70	3	150	—	—	20	2	0,7	200	1	20	—	30
5	Quartz-feldspar rocks	5	5	0,01	0,005	<0,005	—	7	1,5	3	10	20	5	—	—	100	7	3	—	—	20	30	10
4	Quartz-sericite schist	>5	3	0,1	0,15	0,005	30	5	1,5	7	300	7	10	200	150	15	1	1,5	20	1	200	15	—
3	Cataclastic rock	—	0,7	1,5	0,005	0,03	10	5	1,5	3	<10	3	1	—	—	3	0,3	—	—	—	20	—	30
2	Biotite-sericite schist with chlorite	1		0,015	0,7	0,02	500	500	300	1000	50	3	350		—	15	1,5	2	500	—	1000	5	50
1	Amphibolite	—	5	0,01	1	0,15	700	150	150	150	150	3	200	—	—	30	2	0,7	20	—	300	5	100

*The K, Na, Ba, Ti, and Mn composition given in percentages; the rest in g/ton.

been investigated at all in the 1960s and 70s. Only in the 80s did literature references appear about the likelihood of halmyrolytic alterations on the Precambrian ocean floor [29], and the possible development of hydrothermal-sedimentary substances on the Precambrian land surface, resembling weathering horizons [9].

Surface Hypergenesis

Among Precambrian rocks, eluvial deposits and their reworked products have been prominently noted in the literature. The illuvial genesis of caliche was cited. No references have been made at all to ore caprocks, selective solution products of carbonate, sulfate, salt and other deposits.

Problems of the Precambrian Weathering Horizons. Deep-metamorphic rocks, related to Precambrian weathering horizons, have been found in practically all Precambrian shields and shield extensions of the USSR [3, 29]. Several hundred outcrops of "Precambrian weathering horizons" had been noted in professional journals and reports. However, all are thin and their dimensions may not be compared with the thick post-Silurian Phanerozoic kaolinitic and lateritic eluvial deposits (Fig. 1).

Well-known investigators of the Precambrian Shield as early as 1975 have noted that "... no complete profiles of the weathering horizons are known presently" ([1], p. 122). Fifteen years passed since without new results. Does the absence of such horizons mean that they never existed or have they been removed by erosion? Practically all researchers, the works of which we are familiar with, reply that the horizons once existed but have been eroded: "... presently we see only the lower parts of the weathering horizons, not eliminated by erosion. They belong to the illite zone. The processes affected higher horizons more thoroughly, leading to formation of kaolinite and apparently also of free Al-hydroxides" ([2], p. 17, etc.). If the authors are correct, then the Precambrian processes that formed thick weathering horizons, also caused intensive chemical rock alterations (with formation of kaolins, laterites, ochre deposits) and promptly eroded the upper part of the weathering horizon down to the level of the illite zone. It would be strange if such processes would have ceased with the start of the Devonian, exactly at the time when commercial hypergenetic ore deposits accumulated through surficial hypergenetic processes [15, 17].

Assumptions that the Precambrian weathering products as the result of their greater age, have undergone greater erosion than, for instance, Mesozoic horizons, are groundless. Weathering horizons, "imprinted" in Precambrian rock units, were as likely to be preserved as were Mesozoic eluvial deposits.

The term "thick", often used in general characterizations of Precambrian weathering horizons ([24], p. 12, etc.), as a rule disappeared when expert workers started to describe specific sections of the ancient eluvium (works by V. A. Sokolov, K. I. Kheiskanen, A. S. Koryakin, A. D. Dodatko, V. K. Golovenok, M. M. Ipatov, V. M. Chaik, etc.)

The analysis of the substances led to the following conclusions: the presently existing weathering horizons were never thicker than a few meters. The 20-60 m thickness values, found in a number of publications include source rocks, chemically not distinct from the weathering horizons and subjected only to very early stages of alteration. The dark minerals were partially chloritized, the plagioclase minerals became saussuritized and sericitized. Quartz disintegration and other processes took place, characteristic of various later-stage rock alterations (metasomatism, retrograde metamorphism, hydrothermal alterations, etc., ranging to modern subsurface hypergenesis). The weathering horizon itself, if present in the Precambrian, occurs only in a vestigial fashion.

Thus Dodatko and others [4] speak of a section of "metamorphosed plagioclase granite weathering products," c. 100 m thick (Saksagansk region of the Krivoi Rog Basin [Krivbas], Drillhole #8406). The section includes the following sequence: 135.0-112.67 m: plagioclase granites with strongly sericitized plagioclase feldspars; 112.67-107.2 m: quartz-biotite-muscovite schists; 107.2-40.75 m: quartzites. Thick amphibolite beds occur above.

If any part of the section belongs to the "ancient weathering horizon," then it is only the quartz-biotite-muscovite schists. "Relict plagioclase grains, with corroded polysynthetic twins occur rarely in the thin section" ([4], p. 127).

Attempts have been made in other instances to determine the original thickness of weathering horizons by calculating the volume of assumed reworked products. Often significantly younger (Carboniferous or even Mesozoic) eluvium had also been transported over Precambrian surfaces. Thus, "Traces of Late Archean weathering occurs in the Kursk Magnetic Anomaly (KMA) area and the Krivoi Rog Basin, along the rim of the Dnepr-Donets Aulacogen as weathering horizons that extend 800-1000 m into the basement" ([18], p. 164). However, in the cited areas the Precambrian basement is either covered by Lower Carboniferous (KMA) or by Mesozoic-Cenozoic deposits (Krivoi Rog Basin). Carboniferous and Cretaceous spores provided the post-Precambrian ages of the weathering horizons here, in zones of intensive (several hundred meters deep) hypergenesis [12].

In describing specific examples of Precambrian weathering horizons, researchers usually note that presently only metamorphosed products of the zone of decomposition, leaching and illitic alteration occur. To be fair, several workers attempted to recognize the zone of kaolinitization and even of laterization in sections of Precambrian weathering horizons.

One work is very typical in this regard [26]. The authors included "greenish-gray quartz-sericitic rocks into the Proterozoic (Jatulian) weathering horizon, with decomposed granitic texture, often with significant schistosity..., the base of the sequence contains thin sericite flakes" ([26], p. 178). These sericitic schists grade into weathered, fractured granite downward in the section. At most, they are a few meters thick. Despite the fact that the alkali and alkali earth element content totaled c. 10% in the upper part of the layer, considered to be the weathering horizon, the authors attempted to treat the sericite schists as a kaolinite zone. They assumed that the lower sericite schist interval formed through illite metamorphism, while the upper, "kaolinitic" interval after burial of the weathering zone initially was enriched in potassium "due to the impact of alkaline ground waters on the original clayey products of weathering" ([26] p. 183).

Koryakin [10] agreed with Sokolov and Kheiskanen. However, judging from his later publications, he discarded the idea of complex, improbable metamorphic processes and thought that "reconstruction of the original composition of eluvial products of the investigated profiles indicate that their maturity level corresponds to that of illitic zone development, but not to that of the kaolinitic zone" ([29], p. 85). No direct evidence is cited in the literature for the "development of lateritization processes in the Precambrian...as a regional phenomenon" ([23] p. 1153).

Advocates (including Academician N. M. Strakhov) of a widespread distribution of lateritic weathering horizons in early stages of the Earth's evolution base their conclusions either on theories of lithospheric evolution, or utilize indirect, multivariant facts. For example, "the presence of conglomerate rich rocks in reworked metamorphosed weathering horizons of the Patom Mts. and the Kola Peninsula (chlorite, chlorite-mica, mica schists, amphibolites — B. M.) indicate the variety of lateritic weathering horizons in the Precambrian. Their parent rocks, among others, may have included granites ([5], p. 201).

Most supporters of the laterite development theories on Precambrian continents currently only "rehash" their dated assumptions. However, a few careful statements have been made at the end of the 1980s about "local weathering horizons that are very rare in the shield areas and probably are of lateritic composition (low SiO_2 and alkali metal values, high iron content, etc." ([9], p. 22).

Thus, the rock composition, associated with Precambrian weathering horizons differs sharply from the composition of post-Silurian eluvium. Horizons formed through intensive surface leaching of rocks, typical of younger weathering horizons, are essentially absent from the Precambrian. The younger horizons originally consisted of kaolinite, hydroxide minerals and iron. The "Precambrian weathering horizons" are characterized by isochemical metamorphism, usually accompanied by K-metasomatism that essentially distinguishes them from the Post-Silurian eluvium.

Reworked Products of Precambrian Weathering Horizons. In the literature of the 1970s-1980s this term usually meant a complex of very "mature" terrigenous sedimentary rocks and their chemical differentiation products: "high-Al shales, mica-bearing and monomineralic quartzites" [1] (p. 122), "... often large volumes of conglomerate-rich and quartz-bearing beds may be explained by prolonged weathering horizon-forming processes and repeated deposition of unconsolidated material of the weathering horizon residue [29], p. 229). The insignificant thickness of the sericitic-quartzitic schists, in comparison with the zone of the younger primary kaolinitic weathering horizons, as well as the widespread development of quartzites and muscovite quartzites indicate intensive erosion of the older weathering horizons and redeposition of these products ([4], p. 136). If so, a number of researchers undertook the reverse task: "V. Z. Negrutsa subdivided three weathering levels during the Archean—Early Proterozoic interval in central Karelia, in Pre-Summiisk, PreSariola and Pre-Jatulian times. The first two refer not to weathering horizons but their reworked products. Metamorphosed erosion products of chemically highly mature weathering horizons (such have not yet been recognized in Karelia — B. M.) of Precambrian basement rocks that consist essentially of quartz and other weathering-resistant minerals" ([29], p. 98).

The question arises: why should high-Al and quartzitic rocks without fail be erosion products of ancient "chemically highly mature" weathering horizons?

High-Al Rocks. The term *high-Al rocks* generally was and is applied to effusive and, less frequently, to deep, regionally metamorphosed rocks, that when their chemical composition is calculated by Zavaritskii's method, fall in the "Al-saturated" group. But Al-saturation, according to Zavaritskii refers not to a high Al-content in the rock, only to the Al-excess over Ca and alkali elements. "The term, high-Al rocks refers to a rock group that is, in molecular quantities saturated with Al, in contrast with alkali elements and Ca. In these, $\text{Al}_2\text{O}_3 > \text{K}_2\text{O} + \text{Na}_2\text{O} + \text{CaO}$. They include corundum, disthene, viridine, spinel, "pyralspit" group garnets, cordierite, sapphirine, tourmaline, micas, and hypersthene of higher Al content. "The Al content in the cited minerals of these rocks was not limited" ([13], p. 8). Right there, Kulish included the following to the high-Al rock group

in his "not limited" category: "quartzites with sillimanite, garnet and cordierite (quartz content: 80-99%, SiO_2 : > 83%), quartz gneisses with sillimanite, cordierite, and garnet (quartz: 70-80%, SiO_2 : 74-83%)" ([13], p. 61). To eliminate this obvious terminological inconsistency, Golovenok used the term "high-Al rocks" only for rocks that, in his view, formed through alteration from essentially kaolinitic and lateritic weathering horizons. "High-Al sedimentary—metamorphic Precambrian rocks are defined as metamorphosed sedimentary, mostly clayey rocks that contain at least 25% Al with minor or insignificant amounts (usually not more than 3-5%) of alkali and alkali earth elements. The silica ratio is close to or greater than 0.50" ([2], p. 45).

It is significant that there are but a few (including V. K. Golovenok) who subscribe to this definition. The literature on Precambrian weathering horizons usually assigns to high-Al rocks, metamorphosed rocks (schists and gneisses) that contain low-grade metamorphic minerals, including pyrophyllite, diaspore and chloritoid; at higher grades, staurolite, andalusite, cordierite; at the highest levels, disthene and staurolite'1 ([2], p. 105). Without evidence, such rocks uniformly have been defined as metamorphosed kaolinitic clays, "including also admixtures of other clay minerals, primarily illite" ([2], p. 104). Published tables show but a 15-18% Al content in them.

The assignment of all Precambrian quartz—Al metamorphic rocks to products of reworked weathering horizons is highly questionable. Quartz—Al rocks, known as nonsedimentary in origin, are very common. They include diaspore quartzites and quartz diasporites that form during hydrothermal alteration of volcanic rocks (alunitic deposits of the Amur lowlands and Transcarpathia), corundites among secondary Central Caucasian quartzites, etc.

Detailed results of research done by colleagues at VIMS have been published in 1989. These clearly demonstrated the absence of any link between diaspore mineralization and surface hypergenesis in these ancient rocks. Investigating the rock material, they equivocally stated that the "Upper Proterozoic and Paleozoic hydrothermal-metasomatic diaspore mineralization was unrelated to bauxite formation. Therefore, in prospecting, such a mineralization can not be considered a bauxite indicator" ([6], p. 27).

Cherkasov, Luzgin, Mamaev, et al. contemporary workers attributed hydrothermal-metasomatic origins to many quartz-kaolinite (quartz-dickite) rocks.

Quartzite Rocks. Quartzites and muscovite-quartzites are among the most widespread Precambrian quartz rocks; since the Proterozoic, quartz sandstones and conglomerates are predominant.

Most workers relate quartzites to ancient chemogenic sedimentary formations. The discovery of these rocks in the early Precambrian (Early Archean) led certain workers to confirm not only the existence of large water basins at that time but also thick chemically weathered horizons. Their erosion resulted in quartzites. In particular, Salop ([22], p. 60) explained accumulation of "chemically well differentiated rocks, such as quartzites, quartz-sillimanite schists, by the existence of weathering horizons in the Katarchean, formed through chemical disintegration of sialitic rocks." In his view, from this hypothetical weathering horizon did silica enter the proto-ocean.

Wouldn't it be easier to disregard the idea of thick hypothetical Precambrian weathering horizons that nobody has yet actually observed?

It is more likely and demonstrable that hydrothermal-sedimentary lithogenesis, with associated underwater and subsurface lithogeneses were widespread in the Precambrian, especially during the Archean and Proterozoic. Intensive dehydration in the Precambrian of ancient crustal matter produced significant amount of highly mineralized thermal waters on continental surfaces and ocean floor. This resulted in accumulation of large volumes of colloidal matter [20]. The numerous, remarkably stubborn defenders of an "essential similarity between Precambrian and Phanerozoic hypergenesis" assumed an interrelationship between quartzites and processes of intensive chemical rock weathering, supposedly developed synchronously on the Precambrian continents [1, 2, 4, 9, 13, 22, 24, 29, etc.]. For one reason or another, they completely disregard well known facts about the accumulation of thick beds of monosiliceous matter either near volcanic centers (e.g., diatoms and tripoli of Kamchatka and the Far East), in marine basins, forming through microbiological activity (e.g., radiolarian muds in the southern World Ocean; tripoli and opoka in continental seas of the Russian Platform, etc.), or widespread beds of "secondary quartzites" (including diasporites and corundum rocks) in areas of continental volcanism. There are additional examples for formation of monoquartzitic rocks, clearly unrelated to weathering processes (e.g., extensive jasperite beds, certain jasper types, etc.)

The analysis of the genesis of ancient quartzites and their derivatives [16] provides ample evidence to assume that they formed predominantly through silica transport by thermal (but not necessarily juvenile) waters. These percolating waters reached basin floor areas or, directly, continent surfaces. Quartz sands and conglomerates in such instances may have formed through decomposition and erosion of these, essentially hydrothermal-sedimentary quartzites.

Infiltrational Weathering (Illuvial Horizon). Many workers [9, 27, 29, etc.] questioned the formation of caliche-type carbonate weathering horizons in the Precambrian. These usually are thin (1-2 m) carbonate beds (limestones, dolomites) in quartz-sericite sequences. Thus, Sochava and other ([27], p. 1452) dated the following section of a "metamorphosed weathering horizon and the overlying carbonate weathering horizon" Jatulian (Lower Proterozoic).

The Archean plagioclase granites are overlain by schists with relict textural characteristics of granites. In the upper part, the beds contain disseminated dolomite. The carbonate concentration increases upward in the section and gradually replaces the quartz-sericite-chlorite matter. The dolomite beds, interpreted by workers as ancient caliche, are 2-2.5 m thick. The contact between caliche and the overlying metasandstones and metagrelites is sharp, indicating erosional origins. Calcareous beds occur in the terrigenous Jatulian. Occasionally, they include significant amounts of dolomite fragments. This shows that the carbonate weathering horizons, previously more widespread, have undergone erosion. Information on this substance and its comparison with modern (Cenozoic) caliche beds provides the basis for considering carbonate layers at the base of the Lower Proterozoic as caliche.

However, carbonatization is characteristic of numerous younger rock alteration products, in particular in retrograde metamorphism (diaphtoresis), regressive epigenesis (according to L. B. Rukhin) and subsurface hypergenesis. Furthermore, similar substances probably may accumulate in "shallow temporary basins" ([25], p. 39). Products that form by such processes often resemble caliche. Therefore, mineral composition alone is insufficient for confirming illuvial-surficial origins in certain Precambrian carbonates. Specialized geological research is required in this regard (establishment of a continental erosional surfaces, study of redeposition products of caliche, indicators of an arid climate, etc.). The claim, that "Ca concentrated in the middle part of the weathering horizon (the quartz sericite beds — B. M.), while in individual instances it has extensively replaced the rock body" lacks any foundation. Sokolov ([29], p. 94) compared such carbonate rocks with the caliche in modern deserts.

Underwater Hypergenesis (Halmyrolysis)

While theories vary on the formation history of oceans, no one question, that water basins did exist on the Earth's surface at the start of the Proterozoic. It can not be denied either that during early, Archean and Proterozoic stages of the Earth's geological history the thermal regime of the crust was more intensive. Intensive gas and juvenile water flux percolated through the crust, associated with degassing and dehydration of primeval earth matter. Under these conditions the water basin (ocean) floors were chemically and physicochemically extremely active systems. Not only underwater lava flows but all rock complexes of the ocean floor participated in thermal hydration processes. Underwater hypergenesis (halmyrolysis) must have been a dominant feature in near-surface rock alteration during the Precambrian.

Workers who studied the makeup of basalt halmyrolysis products on the modern ocean floor during past decades indicated that significant composition changes have taken place in them. In addition to intensive hydration and almost total loss of Ca, significant decrease in Mg and occasionally in Na content took place. Iron oxidation and sharp increase in the K-, partially also in the Al-content had taken place [11].

In comparing the composition of modern halmyrolytic (palagonitized) products of basites and of "ancient weathering horizons" with basic rocks, all rock-forming elements displayed practically total identities (Table 1). The one standard exception is represented by water in modern halmyrolytic products of basalts. Water obviously was lost during metamorphism, occasionally along with Na that was absorbed from modern sea water.

Naturally, a direct analog is quite risky in certain rock sequences between "ancient weathering horizons" and modern zones of halmyrolysis on the sea floor. However, it is more justified than assuming "essential similarity" between modern and ancient weathering horizons. Such conclusions were based on the conserving nature of ocean floor sediment alteration, associated with steady hydration of lava during its outpouring on the sea floor, and its concurrent, relatively high-temperature leaching, associated with active hypergenetic metasomatism.

It is noteworthy that as early as 1978 E. Dimrot and A. P. Likhtblau discussed the great similarity between Archean iron-enriched crusts on Canadian Shield pillow lavas and Cenozoic halmyrolytic products on New Zealand pillow lavas ([29], p. 81).

Subsurface Hypergenesis. Complex exo- and endogenic energy sources are the distinguishing marks of subsurface hypergenetic processes. Exogenic energy from space (mostly as solar energy) is transformed on the Earth's surface and later transformed inside the lithosphere through different water categories, accumulated in crystal lattices of hypergenic minerals, in colloids, organic compounds, hydrogen sulfide, etc.

Endogenic energy is active in the Earth interior, according to Chukhrov, in utilizing various hypogenic water (elisional, artesian, and partly juvenile-hydrothermal) categories. These waters are mixed in the hypergenetic zone and their rock alteration products bear similarities both to exo- and endogenic substances.

The aggressive character of subsurface waters (in contrast to surface waters) is determined not only and not so much by a high redox potential and low level of mineralization; on the contrary and at times by high mineralization levels (brines), enrichment in petroleum bitumens (low Eh values), high (50-80°C and higher) temperatures, sharp fluctuations between acid and alkali parameters (pH from 1-2 to 10-12), and other characteristics.

The role of essentially hypergenetic ground waters was enormous in the alteration processes of Precambrian rock complexes. A large body of literature deals with the analysis of data and modeling of various processes of artesian water percolation, elisional leaching, retrograde metamorphism, formation of thick zones of acid leaching and other features that caused alteration in ancient beds. Unfortunately, comparison of products of these processes and "ancient weathering horizons," although attracting great interest, has not yet been undertaken.

At the Seminar of the Commission on Formation of Metamorphic Ores the "classic section" of the old weathering horizons between the Mikhailovsk (A2) and Kursk (PR1) Series in the KMA region have been discussed. M. M. Ipatov and I. M. Kostyuk, associates of ILSAN, studied the section in Drillhole #5957; Drillhole #5257 was investigated by us.

Following detailed studies, Ipatov and Kostyuk ([9], p. 61) reached the conclusion that the formations, considered by participants of the Seminar as the "classic Precambrian weathering horizons" actually represent a zone of "postmagmatic alteration of granitoid-type greisens within a granitic massive".

Their mineral content data was completely identical to that of a complex that we have documented, sampled and studied by various methods. The complex occurred not inside granitic intrusion but at the contact between the Mikhailovsk and Kursk Series (Fig. 2, Table 2).

Figure 2 shows two interpretations of the cited section in the drillhole. Defenders of the Precambrian weathering horizon concept (Ya. N. Belentsev, Corresponding Member, USSR Academy of Sciences, A. D. Sadko and I. N. Shchegolev, Doctors of the Geological-Mineralogical Sciences, and others), at the Seminar claimed the existence of a major stratigraphic and angular unconformity at the boundary between the two Series, essentially different types of metamorphism, the probable presence of basal formations at the start of the Early Proterozoic transgressive cycle, the development of sediments at the contact with continental deposits that characterize the variability of the section with frequent facies transitions, as well as the presence of erosional surfaces within the sequence. These and many other indications point to a large-scale unconformity hiatus during which a several km thick sediment sequence had been eroded. Erosional downcutting thus uncovered Archean amphibolites, products of deep regional metamorphism in the Early Proterozoic surface. However, nothing like this has ever been observed in any of the sections that include the Archean-Proterozoic contact. The thick, persistent metamorphic rock units generally can be traced to great distances. While chlorite-biotite schist and amphibolites occur both in the Mikhailovsk and the Kursk Series, amphibolites are more typical of the Archean, and schists more common in the Lower Proterozoic.

The assumption is that the drillholes did not penetrate a weathering horizon but an ancient aquifer, traversed by thermal waters for an extended time period. This theory conforms better with available data. Such features, well known from the literature are zones of acid leaching. This kind of origin is clearly indicated not only by the petrographic composition but also by the surprisingly symmetrical distribution of rocks, altered to different degrees, in relation to the central part of the zone, composed mostly of high-temperature quartz. Significant differences in the rare and trace element content between rocks of this zone and Archean amphibolites (Table 3) indicate the presence of a "paleoaquifer horizon."

In describing the "oldest pre-Kursk weathering horizon", many authors noted a zone of "decomposed amphibolites with amphibole-biotite-chlorite cement" ([19], p. 69). In their view, decomposition occurred much later, when the amphibolites were already exposed in the land surface and the leached Kursk Series rocks had been converted into clays, mud and sand. Subsequently both series were metamorphosed. Amphibolites are found in both series. Why were the decomposed amphibolites not converted into massive rocks again? Because their hydrolysis, "decomposition" by aggressive hot hydrothermal ground waters occurred much later. Certain workers, as A. I. Nikol'skii, related intensive hydrothermal processes to Late Proterozoic granite intrusions.

The other, not less important example is an attempt by certain authors to interpret the basal horizons of the Upper Archean Keiv Formation in the Kola Peninsula as the "Pre-Keivsk weathering horizon".

The Keiv Formation is composed essentially of quartzites, conglomerates, mica-biotite- and high-Al-schists, and amphibolite. Their total thickness is c. 1.5 km. Its lower-portion includes few beds of quartz-kyanite rocks and ores, with a mean kyanite content of 30-40% (Fig. 3).

During the past few decades, researchers of the Kola Peninsula Precambrian regarded the lower portion of the Keiv Formation as a product of more recent, deep acidic leaching. The leaching products reflected the primary composition of the source rocks. Thus, Zhdanov ([7], p. 186) indicated that acid leaching of anorthosites in the Keivsk zone developed an essentially kyanitic rear rock belt. In terms of chemical changes, acid leaching effects are identical to those that exist during weathering horizon development. The process is accompanied by leaching of alkali and alkali earth elements and accumulation of an amphoteric base. The differences between these processes are shown by the varying temperatures and pressures that resulted in essential differences in the mineralogy of the end products and in the makeup of the sections.

The following evidence was cited by Golovenok, Bel'kov, Mirskaya, Negruts and others to claim the existence of a PreKeivsk weathering horizon associated with this section: (1) bleaching of gneisses and gradual transition toward quartzitic rocks, overlain by the Keiv Formation, (2) gradual increase in the muscovite and quartz content in the gneiss unit; (3) garnet in the altered gneiss, in the rutile-zircon heavy mineral fraction; (4) high-Al schists themselves were regarded products of immediate redeposition of a ferrallite zone (lateritic — B. M.) in the Keivsk weathering horizon.

All these features are fully compatible with a deep-metamorphic-metasomatic genesis of the Keiv kyanite schists.

As in the previous example (KMA area), no primary evidence is available either for a continental unconformity surface (strongly dissected relief, coarse-grained sediments, lithofacies variations) and for the development of ancient, thick weathering horizons.

Many of these features, noted in the literature as "Precambrian weathering horizons," instead, may be described as products of various younger subsurface rock alteration categories, here combined by the term, "subsurface hypergenesis."

RECONSTRUCTION OF PRECAMBRIAN HYPERGENESIS

Despite the fragmentary character of our knowledge about the Precambrian evolution of Earth, it may be said that Precambrian hypergenesis was a component of lithogenesis. Subsurface hypergenesis was the most conservative process throughout the Precambrian. Surficial hypergenesis products have undergone the most intensive changes.

Three evolutionary stages were distinguished during Precambrian hypergenesis: Early Archean (4.5-3.5 billion yrs), Late Archean-Early Proterozoic (3.5-1.6 billion yrs) [16].

Early Archean Stage. Conclusions in the literature on Early Archean hypergenesis are contradictory and usually speculative. They are based exclusively on the reconstruction of the original appearance of deep-metamorphic rocks; biotite, garnet-biotite, graphite-sillimanite-cordierite gneiss, crystalline schists, marbles and quartzites. Earlier these rocks may have had a chemical composition, similar to modern surface deposits (clays, silts, sands, etc.). They formed under physicochemical conditions and in media that were similar or even analogous to modern conditions. A certain similarity of the chemical composition with those of the above-cited rocks in certain eluvium types is one of the basic "indicators" of the metamorphosed state of the weathering horizon.

Additionally, the most prevalent assumptions in the literature represented different points of view with regard to the initial appearance and composition of Early Archean hypergenesis. For example: "The high weathering intensity and the thick weathering zones during the Archean and pre-Archean (these have not been described in the literature, only assumed — B. M.) were caused by a particular greenhouse effect, due to interaction between volcanogenic silica and Mg of the sea water at temperatures of at least 80°C ([9], p. 13).

A recent, more radical theory with regard to Archean hypergenesis was presented by Ipatov and Kostyuk ([9], p. 60). As far as quartz-sericite schists ("metamorphosed weathering horizons") are concerned, they assumed that "such rock composition developed during the post-magmatic development stage of hydrothermal deposits and is known as the process of rock beritization" ([9], p. 60). The authors assumed that the pt-conditions that existed on Precambrian continental surfaces favored beritization. There were neither illite, kaoline, nor free Al- and Fe-hydrate minerals. The Precambrian surface hypergenesis (crust formation) was characterized by the direct replacement of magma minerals by a hydrothermal (in part, berizitic) mineral association.

Analysis of data on the evolution of the Earth's crust confirm the present assumption with regard to the Early Archean as a pre-hydrospheric stage of lithospheric stage" ([9], p. 45). High pressures and temperatures on the Earth's surface, a steady flow of gas and gas—liquid emanations from the interior, formed layers of a heavy, aggressive atmosphere, saturated by vapors of various acids. Hypergenesis in such media differed little from conditions that prevailed under deep hydrothermal-metasomatic alteration conditions. Intensive heating of the primeval matter from great depths provided the main energy source for hypergenic (near-surface) processes.

Until now, no creditable evidence existed in the literature on Precambrian continental deposits that may be interpreted as indicating the presence of erosion products of weathering horizons. As far as the existence of marine sedimentary beds is concerned (quartzites, marbles, sillimanite schists, etc.), their development was associated with the transport of enormous amounts of various compounds to the surface by juvenile waters during dehydration of primeval crustal substances in initial stages of their development.

Late Archean-Early Proterozoic Stage. At the start of Late Archean, a geologically fast evolution of basic hypergenetic, atmospheric, and hydrospheric factors resulted in formation of the biosphere. Large marine basins formed and the water-land-atmosphere interfaces became sharply defined. The products of all three hypergenetic categories have been defined by their physical properties. Weathering horizons evolved on the continents, as the interface zone for interaction between rocks, exposed on the land surface, and the atmosphere. Halmyrolysis processes developed on the ocean floor, essentially involving volcanic rocks. Rock leaching: "subsurface hypergenesis," occurred in the shallow subsurface.

The following circumstances regulated surface hypergenesis: (1) the absence of plant matter on the continents; (2) higher carbon dioxide partial pressure in the atmosphere and the correspondingly more acidic composition of surface waters than in the Phanerozoic; (3) intensive deep thermal water discharge, resulting in higher ground water temperatures and activation of hydrolysis processes.

The impact of vegetation on modern surface hypergenesis has been studied by a large number of workers. Their results show, first, that absence of vegetation results in a general sediment deficit and climate zonation; precipitation in coastal zones, aridity in continental interiors. This, in turn, results in great atmospheric pressure fluctuations and corresponding increases in wind erosion intensity. Wind erosion apparently was a prime factor in the Precambrian denudation of continental interiors. Second, vegetation sharply reduces the surface discharge coefficient and in certain instances alters it by infiltration of water into the ground and its subsurface migration. In the absence of vegetation in the Precambrian, the waters were practically uninhibited on the surface. These temporary streams possessed no established channels with stable flow. They drained in areas of accumulation, causing localized, mostly mechanical erosion.

High temperatures and thermal water inflow in the shallow subsurface of the Earth led to a significant increase in the absolute humidity values of the air in the soil. This intensified hydrolysis of the silicates.

Surface hydrolysis products were not stabilized on the land surface during the Precambrian. Intermittent streams after their formation were transported to areas of sediment accumulation. No major accumulations of weathering horizons formed in the erosion zones.

Surface Hypergenesis during the Late Archean-Early Proterozoic thus was associated with intensive hydrolysis of Al-silicates and local erosion of its products to water basins. No "thick weathering horizons, formed by intensive chemical weathering" occur in the Precambrian. This conclusion, reached as early as the mid-1970s, has been confirmed by M. M. Ipatov, I. M. Kostyuk, M. I. Graizer, Kh. G. Il'inskaya and numerous other students of Precambrian weathering horizons [9, 29, etc.].

All changes in eroded rocks occurred mostly during sediment transport to basins in the diagenetic stage. Under dry conditions, surficial carbonate (caliche) and silica (silcrete) crusts may form in continental interiors through infiltration processes under specific conditions.

Quartz-sericite and quartz-sericite-chlorite rocks occasionally preserved relict structures of underlying source rocks and formed thin, clearly defined layers. They were described by numerous workers as "Precambrian weathering horizons." After detailed observation, in most instances these may be identified instead as products of halmyrolysis or acid leaching. In certain instances, a surficial hypergenic origin remains a possibility.

A close study of the section is required to correctly establish eluvial origins of certain substances. Investigation of the composition and structure of assumed weathering horizon profiles, the facies and petrographic composition of overlying deposits, and the study of the erosional surface relief must be undertaken.

Late Proterozoic. The nature of hypergenesis did not change during the Late Proterozoic. Due to the absence of land vegetation, primitive desertlike wastelands continued to dominate the land areas. High atmospheric pressures and temperatures, the aggressive chemistry of the atmosphere and rainfall led to very intensive hydrolysis of rocks, especially those that were enriched in alkali and alkali earth elements. While no commercial kaolinite deposits developed, under favorable conditions minerals of the kaolinite group may have formed during rock weathering. Under these circumstances processes of natural Al and Si separation, followed by laterite blanket development, apparently could not have taken place. Al may have been leached out of rocks only by acidic and high-acidic ground waters. In zones where these waters percolate upward (usually in the nearshore zones of oceans), lenticular and bedded silicate diaspore developed, the type known from the Alagul' Deposit, SW Mongolia [17].

During hypergenesis iron accumulated exclusively in water basins throughout the entire Precambrian. Iron reached the basins along with silica from deep subbottom zones. No exogenic geochemical barrier may be envisaged that would have forced iron only to settle. This fact unambiguously and negatively answers the question whether rich ($\text{Fe} > 60\%$) iron ore deposits may occur in Precambrian ore deposits. Such ores could have formed only in mature weathering horizons starting in the Middle Paleozoic.

That intensive chemical alteration of the weathering horizons was impossible in the Precambrian is also explained by the total absence of hypergenic Ni and Co-Ni ores in Precambrian deposits. Such deposits were widespread in the Mesozoic and Cenozoic.

Precambrian hypergenic mineralization was characterized by formation of deposits, almost exclusively associated with subsurface and partially underwater hypergenesis. This led to development of such specific formations, typical of old ores, as the Au—U, Cu—Co, and Au—polymetallic deposits; also a large group of hydrothermal-karst deposits of various mineral resources.

In conclusion, summarized research results of the past years allow critical rejection of the concept of an "essential similarity between geochemical hypergenetic factors throughout the known geological history of Earth." Search for analogs, for an essential similarity between Precambrian and Phanerozoic hypergenesis, leads to a blind alley. One should search for differences between the substances and analyze them. This approach would formulate the principles of rock formation and eventually establish the ability to scientifically identify mineral resource locations.

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PALEOGENE LITHOLOGY OF VOLCANOGENIC–SEDIMENTARY SEDIMENTS OF THE GOVEN PENINSULA (SOUTHERN KORYAK PLATEAU)

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Two structural facies zones are established in the Goven Peninsula which are separated in the present structure by the Primorye overthrust line. The sediments of the northwestern zone are examined in detail. New age determinations based on nanoplankton are reported. A genetic classification is given of gravity flow sediments, which connects their accumulation with the environments of the various portions of the submarine alluvial cones. The sediments of the northwestern zone may be related to the forearc basin.

The Goven Peninsula in the southern Koryak plateau is of interest for reconstructing the history of the northwestern fragment of the continental surroundings of the Komandorskiye basin and for understanding the sedimentation processes with active participation of volcanic activity, in the ocean-to-continent transition zone. The sediments of the peninsula may be related to the formations of Karagin Island, located to the southwest. These groups of formation are combined within the Goven-Karagin block [1]. Some investigators believe that the Paleogene of the Goven-Karagin block was overthrust upon the Upper Cretaceous of the Vatyn suite. They classify these formations as belonging to the Alyutor block, which occupies the central part of the Koryak plateau [6].

As a result of mapping and detailed lithologic investigations, the author, working with A. D. Kazimirov and G. V. Polunin (staff members of the Laboratory of the Tectonics of Oceans and Transient Zones of the Geological Institute of the USSR Academy of Sciences) and L. B. Afanas'ev and N. T. Vorogushin (Kamchatka Geological Association), established in the Goven Peninsula two large structural facies zones: a northwestern zone (comprised of Pylgin ridge, the Atavropel mountains, and their northwestern offshoots), and a southwestern zone (enclosed between the Primorye overthrust line and the coastline of Olyutor Bay on the Bering Sea) (see Fig. 1).

Structural, lithologic, and paleontologic data indicate significant differences between the sediments of these two zones, which have been conventionally combined into one stratigraphic section (by Z. A. Abdrakhimov, A. A. Kolyada, et al.). The basalt, andesite-basalt, and andesite streams are associated exclusively with the northwestern zone (Pylgin ridge proper), where they are interstratified with a compositionally monotonic volcanomictic flysch series. The microfauna in these sediments is represented by very few species. The sedimentary formations of the southwestern zone are much more diverse in composition and are similar to formations in the southeast of Karagin Island. The flyschoid strata are intensely budinized. Individual beds and series are enclosed in an intensely cleaved argillite matrix. The pebbles, often found in the matrix/include sandstone, argillite, and basalts. At the extreme southern end of the peninsula a silicic facies appears within this zone. Thin (a few meters) olistostromes are also fairly common. Eocene microfauna is abundant in sediments of the southwestern zone. According to Serova [4, 5], this is evidence that, by formation time and conditions, these sediments are related to strata of eastern Karagin Island. The Primorye overthrust is interpreted as a tectonic combination of the structural-facies zone. The correlation of these zones should be a subject of a separate study. The present paper is concerned with volcanogenic–sedimentary formations of the northwestern zone.

Three strata isolated in the northwestern zone appear to be elements of a single stratigraphic sequence. The lower position is occupied by a volcanogenic pyroclastic bed (in early schemes, designated by the Ivtygin and Inetivayam suites – A. A. Kolyada), and its base consists of coarsely interstratified tuffoconglomerates and tuff gritstones. In the middle portion, pillow

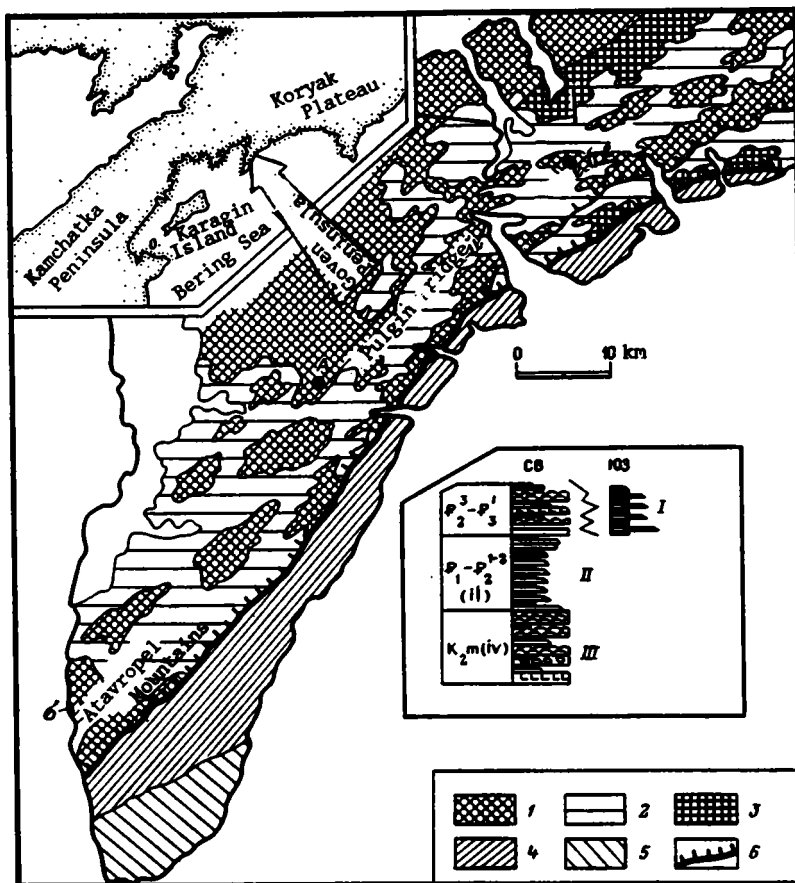


Fig. 1. Schematic geologic map of the Goven Peninsula and a scheme of the stratigraphic correlation of the strata: 1-3) strata of the northwestern structural facies zone (1 — volcanogenic (I), 2 — sedimentary (II), 3 — volcanogenic-pyroclastic (III)); 4, 5 — strata of the south-eastern structural-facies zone (4 — terrigenous, 5 — silicic-terrigenous); 6 — littoral overthrust; A, B — position of lithologic facies sections shown in Figs. 3 and 4, respectively; σ = hyper-basite body.

basalts are common and dominate the sedimentary material in the sections. The top portion is mainly composed of volcanomictic sandstone and silicite, which include isolated basalt flows. The presence of tuffosilicites is typical of this bed. Thin silica inter-layers and tuffogenic—siliceous—carbonaceous concretions are less frequent. The stratum is 600-1000 m thick.

The sedimentary bed (the Ilyyn suite in schemes constructed by Abdrakhimov, Kolyada, et al.), which occurs conformably on the volcanogenic—pyroclastic beds, displays a variety of facies. It consists of volcanomictic sandstones, siltstones, and argillites. The stratum is 300-500 m thick.

The top volcanogenic bed is characterized by extreme facies variability, especially along the course of the zone. In particular, basalt, lava breccias, and coarse tuffs are known only in the northern part of the zone. They disappear completely in the southern parts at the latitude of the southern deep bay. The transition from the underlying bed is gradual: from individual tuff layers to virtually total domination of lava covers and related hyaloclastites and tephroids. The absence of vulcanites in the southern part of the zone makes it extremely difficult to separate the monotonic flyschoids of the sedimentary and volcanogenic beds. For this reason, micropaleontologic investigations, especially studies of nanoplankton, are crucial. The stratum varies from 200 to 600 m, increasing from the south to the north.

Within the northwestern zone, the dating of sediments and stratigraphic separation of beds is difficult, because they are virtually devoid of macrofossils, they vary in facies, are dislocated, and lack clear marker horizons. The dating of the lower volcanogenic pyroclastic bed is based on two finds by Kolyada in 1979 of costate inoceram. The fragments are confined to coarse pyroclastic streams. For the paleontologist T. G. Kalishevich, this is not a definitive indication of a Late Cretaceous age, because fragments of broken inoceram shells are found in Late Cretaceous—Miocene rocks after redeposition by later flows. In

absence of other age data, and considering its structural position, this bed is dated provisionally as Late Cretaceous (Late Maastrichtian).

At the bottom of the sedimentary bed, *Bathysiphon* cf. *Nauss Cyclamina* sp. indet have been found in argillites from the author's collection. According to Serova [4], this corresponds to the Maastricht-Danian boundary deposits. Serova has determined that planktonic foraminifers from the sedimentary bed near the Atavropel mountains, represented by *Globigerina nana* Chalilov, *Acarinina acarinata* Sulb., *A. primitiva* Finlay, and *Globigerina elongata* Glaessner, are of Late Paleocene age. The bed as a whole probably covers a wider age range.

In specimens 90, 251, 256, 258, 264, 271, 274, and 286 (Geological Institute collection), a nanoplankton association has been isolated from the volcanogenic bed. It is represented by the following species: *Coccolithus pelagicus*, *C. formosus* Dictiococcites bisectus, *D. Scrippsae*, *Discoaster deflandrei*, *D. nodifer*, *Reticulofenestra Umbilica*, *R. hillae*, *Cycligargolithus floridanus*, *C. abisectus*, and *Sphenolithus moriformis*.

According to E. A. Shcherbinina (Institute of Lithology and Sedimentology, USSR Academy of Sciences), these species cover a wide stratigraphic interval from the Upper Middle Eocene (*Reticulofenestra umbilica* zone) to the Lower Oligocene. It is a typical deep-sea complex of interoceanic rises. The abundance of fine *Reticulofenestra*, which cannot be diagnosed under a light microscope, endows the set with an appearance reminiscent of formations found in high latitudes in the South Atlantic. In one specimen (258), *Sphenolithus radians*, typical of the Upper Middle Eocene, has been found. These datings link the beginning of formation of the volcanic strata and manifestations of the second stage of volcanic activity to the Upper Middle Eocene, although generally the rocks of this bed can represent a wide age interval.

As is clear from the preceding data, the stratigraphic separation of these sediments is far from perfect. Even so, a revision of the conventional geologic interpretation of the structure of this peninsula is urged by the new data, that contradict the previous theories.

In the past, survey geologists (Kolyada et al.) identified two suites in this peninsula: the Upper Cretaceous–Paleocene Ivtygin suite (and provisionally, the underlying Inetivayam suite), and the Eocene–Oligocene Il'pin suite. All the volcanic formations were referred to the Ivtygin and Inetivayam suites. When volcanites were found at the top of sections, this was attributed to charriage onto sedimentary formations of the Il'pin suite. This view is supported by the fact that normal stratigraphic transitions between beds occur next to disrupted areas with effective discontinuity surfaces formed by intensive folding deformations due to the different mechanical competence of rocks. Contradictions in the theory postulating that all volcanic rocks belong to the same bed came to light in 1969, when Serova discovered specimens from the Ivtygin suite guiding plankton foraminifers of the *Globigerina elongata* Glaessner species, which in her opinion ruled out a Cretaceous dating of these formations. The rock material for these determinations was collected in the sections of the upper volcanogenic bed as identified by us. The author's data on the nanoplankton set referred to the Upper Middle Eocene from the sections of "typical Cretaceous charriages" confirm that the Ivtygin suite should be subdivided into two beds, with sediments reflecting two stages of volcanic activity: the Late Cretaceous and the Upper Middle Eocene.

Though combining the sediments of the northwestern and southwestern structural facies zones into a single stratigraphic section appears unjustified, it is probably a consequence of the classical view assigning a primary role to vertical motion, while overlooking the tectonic combination of initially distant sedimentary basins or portions of a basin as a result of horizontal movements typical for ocean-to-continent transition zones.

The new data call for dropping the term "suite" in favor of a looser designation — "stratum." We discuss the sediments of the middle sedimentary and upper volcanogenic strata, which are convenient for an exhaustive description of flysch accumulation and the relationship between volcanic activity and sedimentation.

Material Composition of Sandstones and Clastic Minerals

By their material composition, sandstones are classified as volcanomictic graywackes. They consist mainly of effusive rock fragments — andesites and andesite-basalts, volcanic glass and plagioclase (andesite and labrador). On Shutov's diagram [7], sandstones are grouped along the side L–F, which is characteristic of volcanic material not diluted with a terrigenous admixture. On Dickinson's diagram [9] of provenance source types, sandstones tend toward the field of slightly eroded island arcs. On Menard's diagram [12] for basin environment types, sandstones are conjugated with the fields of intraoceanic island arcs, partly occupying the fields of pre- and transarc basins.

Monoclinic pyroxene from vulcanites accompanying sedimentation is the dominant mineral (80-98%) of the heavy fraction. Ilmenite, chromite, and magnetite are present in smaller amounts (1-10%). While generally the mafic components of the heavy facies of Pylgin ridge and Atavropel mountain sandstones are similar, the contents of sialic components in them are different. In particular, rutile, anatase, zircon, apatite, and barite contents in Pylgin ridge sandstones rarely exceed a total of 10%. In Atavropel mountain sandstones, sialic minerals are represented by (%): zircon (3.8-58.6), apatite (1.2-60.1), garnet (0.1-14.8), corundum (1.0-2.2), and tourmaline (1.2-1.9) — totaling from 15 to 65%. The presence of corundum and tourmaline many indicate a metamorphic source synchronous to sedimentation. The chlorite-hydromica association of clay minerals in Atavropel mountain sediments appears to be more mature than its counterpart in Pylgin ridge, where it contains smectite.

The true effusive formations of the upper volcanogenic series are associated with the fields of the basalt-andesite and basalt-trachyandesite series on Sabin's diagram [16]. Generally, volcanogenic sedimentary sediments of the northwestern structural-facies zone of the Goven Peninsula can be classified as products of destruction of a mature slightly eroded island arc.

A Notion of the Genesis of the Sediments and Classification Principles

Volcanomictic gritstone, siltstone, and argillite of the two series display well-developed primary structures, which, in combination, favor the hypothesis of deposition of sedimentary matter from gravity streams. Observable exposures are represented by fragments of a continuous sedimentary series formed by underwater gravity processes. These processes begin by entraining large masses of unlithified sediments, resulting in differentiation in the course of transport by variable-density currents. Gravity flow sediments have a specified set of genetic characteristics, reflecting mainly the methods of their transport and accumulation and allowing us to reconstruct their position in the lateral series.

The classification of sediments is based on isolating genetic units of different rank. This is convenient, when classifying and correlating monotonic and paleontologically "sound" strata.

For the primary genetic unit we take a rhythm, which consists of a specific sequence of different lithologic rock types (beds) with characteristic primary structures and thicknesses. A rhythm is formed by a single sedimentologic event and reflects the differentiation of sedimentary matter during the disappearance of a gravity flow. It is an elementary unit of an autocyclic sedimentary process, which allows us to judge about its nature and dynamics.

A set of similar rhythms constitutes a packet, which occupies a considerable space—time volume and can be expressed as a mappable geologic body. A packet combines sediments accumulated in similar conditions during several sedimentologic events and reflects the facies of the deposits, i.e., their landscape associations in a continuous lateral series of gravity flows.

Sediments of different packets succeeding one another in the section constitute complexes of higher rank: cycles which reflect the succession of facies, i.e., sedimentation conditions, and allow one to reconstruct the direction of sedimentation processes.

Within each rank, an interpretive (formational) concept corresponds to the genetic term. It is desirable to describe the sediments in terms of the genetic series, while discussing the sedimentation settings and environments in terms of the formation series, so as to protect the data from interpretive subjectiveness. The following description and interpretation of the volcano-genic—sedimentary sediments of the Goven Peninsula observe these principles.

Description of Packets

Packets of the first type. These combine sediments from conglomerate to sand grain sizes, with a predominance of grit-stone-large-sand units and exotic inclusions and layers of boulder and silt material. The formative beds vary in thickness from 1 to 6 m. The rhythmicity is masked by gradational sorting of sedimentary matter within the layer, which, by virtue of its complex structure, represents a rhythm. Gradation is seen as characteristic of a specific variety of rhythmic pattern. Gradational sorting is typical of sedimentary structures. Simple normal (Fig. 2, 1.1), pendular (see Fig. 2, 1.2), and multiple pendular gradation forms are distinguished. A gradual transition from gradational to massive structure or to a coarse oblique structure is often observed (see Fig. 2, 1.3). Gradational sorting is not usually expressed in an ideal sequence with a lutite roof (after Kuenen [11]) as a result of truncation by a later stream. In nongradational layers, rhythmicity is more manifest as a steplike succession of grain sizes within a bed (see Fig. 2, 1.5, 1.6). In nongradational beds, structures are mostly massive, but coarse and fine oblique stratifications may occur in the lower (see Fig. 2, 1.5) and upper (see Fig. 2, 1.6) portions of a rhythm. In rhythms of a massive structure, areas with a gradational pattern may occur (see Fig. 2, 1.4), but usually they are not developed and are represented by strings of siltstone—sandstone pebbles oriented with their long axes along the course. The presence of chains of coarser material

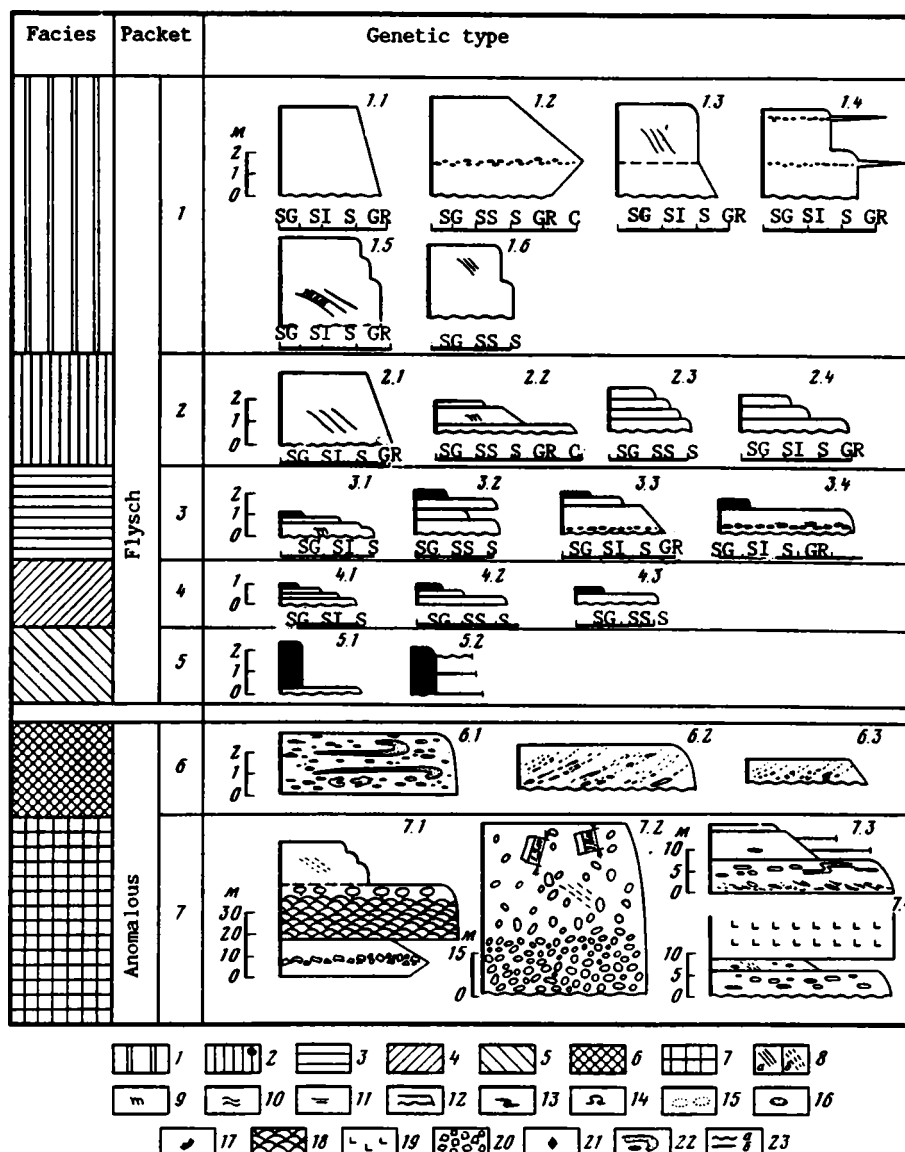


Fig. 2. Genetic types and facies of the paleogene of the Goven Peninsula: 1-7) facies (1, 2 — principal and secondary distributive channels, respectively, 3 — wash-up bars, 4 — vanes, 5 — peripheral parts of cones, tails, and interchannel depressions, 6 — landslides, 7 — volcanic); 8-11) stratification types (8 — coarse large oblique (a — distinct, b — blurred), 9 — fine oblique, 10 — parallel-fibrous, 11 — horizontal); 12) gradation; 13, 14) thixotropic structures (13 — landslide, 14 — convoluted); 15) bioturbation; 16) carbonate nodules; 17) shells; 18, 19) lavas (18 — spherical, 19 — massive); 20) lava breccia; 21) recrystallized radiolaria; 22) fragments of sandy (points) and clayey (black) layers; 23) contact type (a — erosional, b — distinct).

in the rhythms is typical for a packet and is usually associated with gradational rhythms. The position of these chains within a rhythm may be arbitrary, but they tends to be at the base. Pebble size along the long axis ranges from 1.5 to 3.0 and less often 5-7 cm. The composition always corresponds to that of the rhythm. Sometimes landslide structures are discernible in larger chains, but on the whole they are not typical. Differently oriented materials of a large lump size (up to boulders) are sometimes associated with them. No fossils remains have been found. Packets display an irregular lenslike (channel) structure. The longitudinal dimensions are incomparably greater than the transverse size. The relation to the underlying packet is always through an erosional area. The thickness of packets (Fig. 3) attains tens of meters (usually 30-40 m).

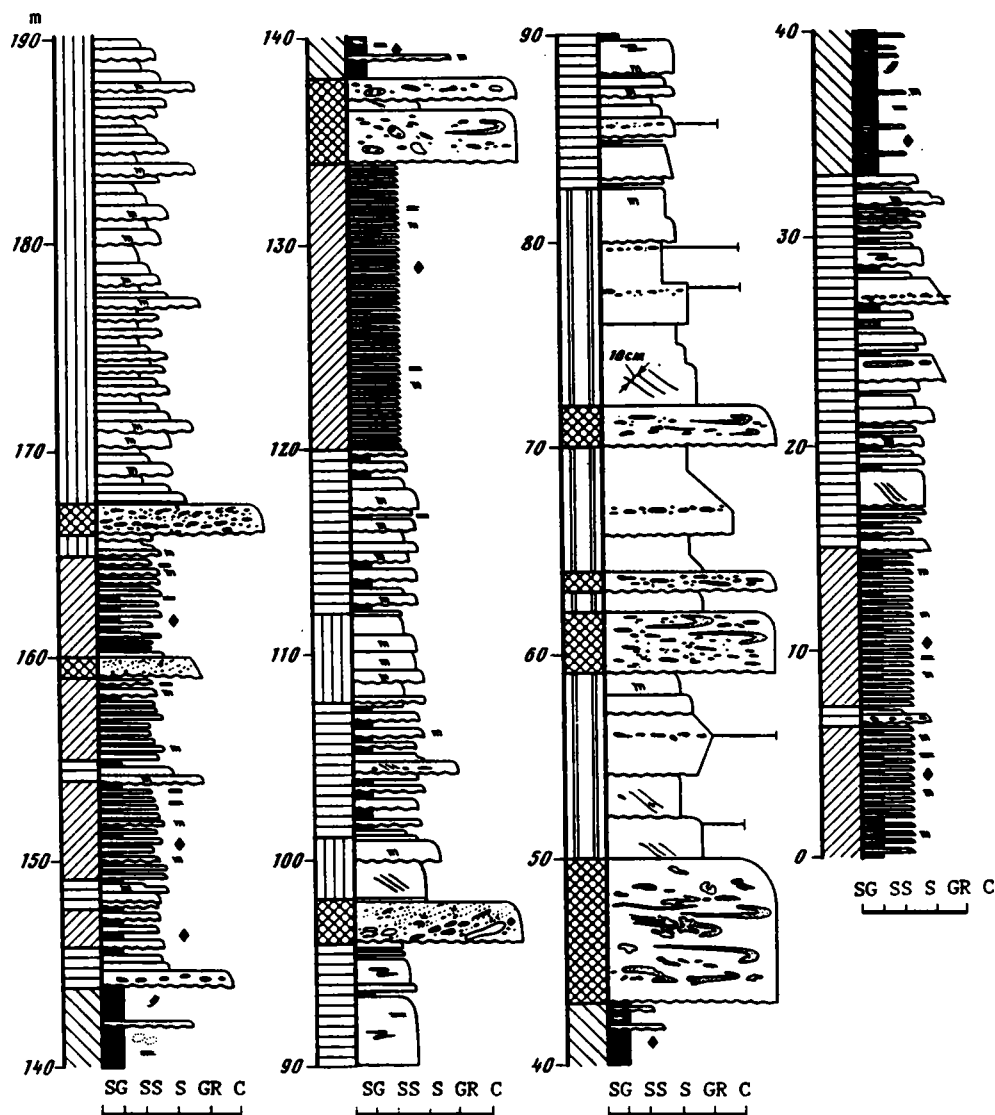


Fig. 3. Lithologic-facies section at the upper reaches of the Gyrgol River under the Ev'vayam Glacier.
Notations as in Fig. 2.

The unequally expressed stratification reflects variations in the hydrodynamic environment during transportation and deposition. In flows that formed normally gradational beds (see Fig. 2, 1.1), sorting was performed because fragments could move relative to each other, indicating a relatively low concentration in the stream and a longer transport as compared with non-gradational and inversely gradational beds [17]. The absence of stratification indicates fast deposition of material from the flow when the time was too short to drag detrital grains. Most gradational beds were presumably formed from high-density turbidity or low-density grain flows. Pendular and multiple pendular gradations reflect the inhomogeneity of the flow in terms of density and transport pattern, and possibly also pulsational fluctuations [17]. The presence of finer material at the base of gradation beds has been studied experimentally. It is attributed to the capacity of finer grains to drown amid larger grains and push them to the surface. Such a mechanism is known as kinetic sifting [13]. Inverse gradation generally indicates intensive transport, reflecting a powerful mechanism triggering the flow and steep slopes, and thus proximity to the source [18]. Chains of pebbles oriented parallel to bedding may be due to flow portions with inertial-granular (fluxoturbidity) transport.

Nongradational beds (Fig. 2, 1.4-1.6) can reflect: 1) deposition from grain flows where clastic matter could not move freely because of a high concentration; 2) complete mixing of material within the bed; or 3) mixing and fluidization of material within the bed with destruction of all structures [10]. The existence of complex-structured packets, transitions between granulometrically different sediments, and different scales of oblique stratification are evidence of transformations in the flow volume

and intensive deposition dynamics. Generally, packets of the first type reflect an intensive hydrodynamic environment and are certainly associated with proximal parts of the sedimentary basin.

Packets of the second type. These include sediments ranging from gritstone to siltstone grain sizes. The layers can be a few meters thick, usually not more than 1 m. The rhythmicity of layering is distinct. A rhythm consists of several layers with clear boundaries. In each successive layer up the section, the grain size is smaller. A rhythm is usually not thicker than 1.5-2.0 m. Gradation in the beds is usually combined with coarse (see Fig. 2, 2.1) or fine (see Fig. 2, 2.2) oblique stratification typical of this type of packet. Nongradational layers are usually massive, but can include differently expressed small-scale oblique stratification. Three- and four-component rhythms (see Fig. 2, 2.3, 2.4) consisting of layers of similar thickness are typical of packets. Chains of elongate pebbles appear quite rarely. No slump structures are seen. Microfaunal remains may be associated with siltstone layers. Lenslike transverse cross sections are smaller than in packets of the first type. The longitudinal size is many times the transverse dimension, and gradual transitions to packets of the first type are observed, while contacts with other sediments are sharply erosional. The packets range in thickness from 10 to 20 m (see Fig. 3).

The combination of gradational and different-sized oblique stratification indicates a slowdown of the gravity flow. Beds are formed in connection with the action of highly concentrated turbidity flows with combinations of gravity deposition of fragments from suspension and bottom dragging. Such conditions can indicate a greater distance from the source and gentler slopes [18] compared to packets of the first type. Nongradational beds were formed in conditions discussed for packets of the first type. Similarities in transportation and deposition mechanisms, gradual transitions, and thinner rhythms, compared with packets of the first type, suggest that these formations accumulated in more distal areas of the sedimentologic object or in landscape elements with less intense hydrodynamics.

Packets of the Third Type. These sediments form a wide grain size spectrum from conglomerate to argillite dimensions, with a dominance of sand grains. The layer thickness rarely exceeds 1 m, usually a few decimeters, and less often centimeters (in argillite beds). A packet consists of layers of various thicknesses and different grain sizes within a rhythm. No order is seen in the succession of rhythms. In layers of certain rhythms a steplike gradation, common in packets of the first type, is combined with fine oblique diagonal stratification (see Fig. 2, 3.1). However, rhythms with massive structures of constituent layers are predominant.

A rhythm may include several components in different combinations (see Fig. 2, 3.2), but multicomponent rhythms are atypical. Gradational layers are not rare but do not form series. Chains of elongate pebbles are confined to the lower portions of gradational layers (see Fig. 2, 3.3) or layers with large grain sizes (see Fig. 2, 3.4). Landslide structures are frequent. No fossil remains have been found. A rhythm is not necessarily completed by an argillite layer, but generally the presence of argillite covers is typical of these packets. The packets display erosional contacts with adjacent units, and tend to occur close to packets of the first and second types. Transverse size is much smaller than the longitudinal dimension (tens of meters). The thickness of packets (Fig. 3) is expressed by a double-digit value (usually 5-15 m).

The different stratification of sediments, the weak rhythmicity, and the lutite covers in rhythms are indicative of an unstable hydrodynamic environment, including periods of high and low activity. Rare separate layers of a normal gradation are similar to such layers in packets of the first and second types. Apparently, they were formed under the influence of low-density granular or high-density turbidity flows with different current conditions [14]. The packets are related paragenetically with the first and second types and are confined to the middle and upper portions of the sedimentologic object.

Packets of the Fourth Type. These packets are sediments, with sand-to-clay grain sizes, and exotic gritstone interlayers. The layer thickness is measured by a few decimeters (usually 10-20 cm). The components of rhythms of similar thickness match within practically the entire packet. The rhythmicity is pronounced. Two- and four-component rhythms (Fig. 2, 4.1-4.3) are clearly separated from one another. Two-component sand-argillite rhythms are predominant. The number of components in a rhythm apparently depends on the transportation distance and the grain size of the material in the flow. Gradation is atypical for sandy layers, which always occur with a clear or erosional contact on the lutite cover of the underlying stream. Sedimentary structures described as c and d elements in the Bouma sequence are common in sandstones [8]. Convolutionary stratification is typical, large inclusions are not found. Microfaunistic remains are confined to argillite interlayers, radiolaria and several forms of planktonic foraminifera have been isolated. The sediments of the packet are widespread laterally and occur in a cloaklike overlap on the neighboring layers. The interstratification and propagation patterns suggest that these are normal-flysch sediments [2]. The packets can be tens of meters thick (see Fig. 3).

Packets of the fifth type. These include sediments of sand—clay grain sizes. The layers vary in thickness. Argillites are clearly predominant. Sandstone and siltstone interlayers are at most 10-15 cm thick, while argillite layers are 0.5-1.5 m thick. The 1-m-thick argillite layers are of an inhomogeneous composition. Microscopic investigations clearly show interlayers of sand—silt material, giving a formal basis for identifying a large number of rhythms. However, for lithologic mapping it is preferable to define rhythms by macroscopic characteristics. Distinct two-component rhythms consisting of microgranular sandstone or siltstone and argillite are well defined (Fig. 2, 5.1). Sand—silt material is often localized in very thin interlayers at the base of a rhythm (Fig. 2, 5.2).

The rhythmicity is of a high degree, and rhythms are grouped into series with similar parameters of constituents, with gradual transitions from one to another. Stratified structures corresponding to the Bouma elements c and d [8] are frequent. Argillites are of a fine horizontal stratification. Microfaunistic remains are found in argillite layers, less often in siltstone layers. Small-scale landslide structures and convolutionary stratification in sandstones are frequent. The sediments of these packets are developed laterally on a large area, with a cloaklike overlap on adjacent formations. A packet can be several hundred meters thick. The interstratification and occurrence patterns suggest that these are distal or subflysch entities [2].

The elements of the Bouma sequence in the packets of the fourth and fifth types suggest that they were formed from low-density turbidity flows. Dragging played an important role in the deposition process, as evidenced by the well-shaped thin oblique stratification in sandstones. Numerous mechanoglyphs on the base of a sandstone layer also indicate a slow current regime [14]. The landslide structures and convolute stratification are probably associated with thixotropic remobilization of redeposited sediment [15]. Packets of the fourth and fifth types clearly tend to peripheral parts of the sedimentologic object or local depressions (Fig. 3).

Packets of the sixth type. These are always sharply distinguished from adjacent deposits. They include materials with a wide range of grain sizes, including fragments of layers and sandy boulders. The composition of sediments matches that of neighboring packets, but single, obviously exotic fragments (such as andesite-basalt boulders) sometimes appear. No rhythmicity is observed. The stratification is usually expressed in the orientation of elongate pebbles and shreds of layers more or less parallel to the bedding (Fig. 2, 6.1). However, variously expressed coarse and rough oblique stratification can also be found, with a certain improvement of sorting from the base to the roof of the oblique thin layers (see Fig. 2, 6.2). A combination of gradational pattern and coarse oblique stratification is found in some locations (Fig. 2, 6.3). The sediments in packets of the sixth type are associated with every type of packet described above (Fig. 3). The contacts with the surrounding formations are always distinct. The bottom contact usually, but not always, is erosional. The packets can be up to 10 m thick, but usually thinner. The lateral extent is relatively small, probably several tens (hundreds?) of meters.

The formation of packets is obviously associated with a thixotropic redeposition of unconsolidated sediment [15] and can be the result of overloading on a slope, a hydrodynamic shock at the discharge of a neighboring stream, the imposition of a tectonic stress, volcanic activity, or a combination of these factors.

Packets of the Seventh Type. They include sediments genetically connected with submarine andesite-basalt effusions: lavas, lava breccia, hyaloclastites, and tephroids. No rhythmicity is seen. The sediments are diverse, but always represent an element of a common lava effusion process.

The packets often reflect a pulseline multistadial pattern and carry a fairly comprehensive chronicle of the volcanic event. In packets with complex structure (Fig. 2, 7.1), psephitic tuffs with pendular or inverse gradation and lava breccias traced by mobile front streams underlie spherical lavas, which form distinct streams. Gradually, they give way to separate pillows floating in a tephroid matrix. The amount of matrix grows up the section, while spherical inclusions gradually disappear. The steplike gradational tuffs at the top of the packet often display a blurred coarse oblique stratification. The complex packets can be tens of meters thick.

Landslide strata and even volcanogenic olistostromes are quite frequent. Packets with distinct lower erosional surfaces of a complex structure are typical. They are thick and include spherical monocomponent fragments of plagioporphyritic basalts floating in the hyaloclastitic tephroid matrix at the bottom of the packet (Fig. 2, 7.2). Up the section, the spacings between fragments increase, and the composition becomes polycomponent: besides plagioporphyritic basalts, amygdaloid and aphyric basalts appear. At the top of a packet, despite the general decrease in fragment size, single olistoplaques of aphyritic columnar basalts are found, which may be as large as 1-1.5 m. Areas of coarse oblique stratification are documented within packets. The olistostromes can be up to 50 m thick. Smaller landslide formations are more frequent (Fig. 2, 7.3-7.4). They often exhibit structures characteristic of packets of the sixth type, with a peculiar combination of several structural types gradational, coarse, oblique stratification, etc. which brings them close to sediments of high-density pastelike streams.

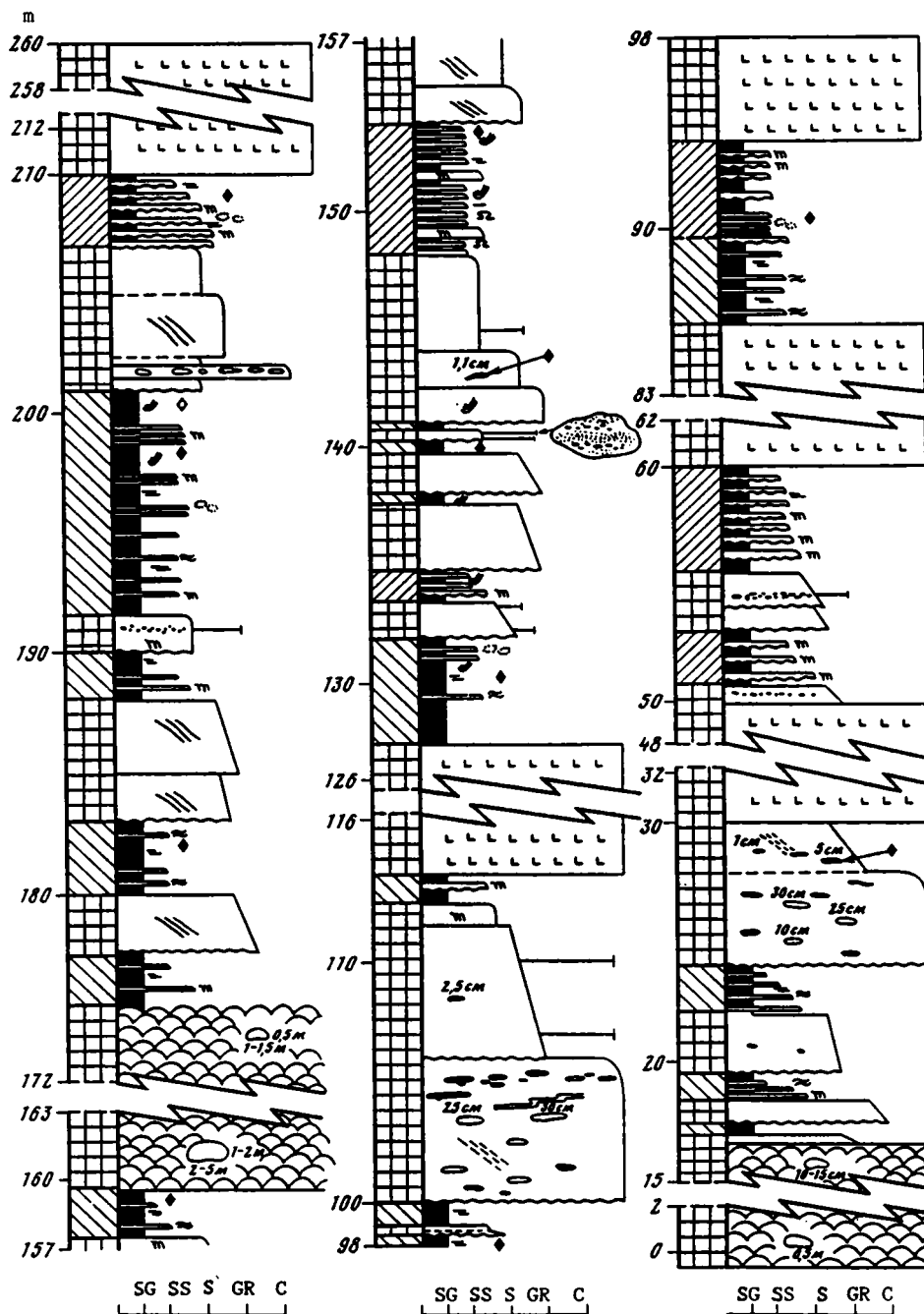


Fig. 4. Lithologic-facies section at the left bank of the Porog River (Usatavayam River basin).
Notations as in Fig. 2.

Thick (up to 20 m) steplike gradational tephroid sediments of a massive structure are frequent. At the base, they usually contain landslide structures indicative of the coherence of the entire body, which survived as a single bed. Its base was deformed early on with accumulation of the roof and overloading by its own weight and inertial action.

Packets of the seventh type. These can be associated with all other types. The occurrence frequency of these packets can be quite high (Fig. 4), representing the environment where volcanic processes determined the deposition patterns in the adjacent portions of the sedimentary basin near volcanic centers. In distal areas, accumulations of synchronous sediments would display only minor reverberations of the volcanic event in the form of single highly mobile tephra flows. The material of packets of the seventh type consists of a mixture of lava flows and surrounding sediments eroded in the course of the motion. The lateral extent of packets

is inconsistent, but their length is much greater than their width. Their thickness varies in a broad range from a few tens of meters to single-digit values.

Interpretation

The presence of specific sedimentary structures of gravity flows, the build of the packets, and their proportions in sections and in plan can be described consistently in the framework of a model of a deep-sea alluvial cone at the base of a paleoslope expressed geomorphologically as an erosional rolling relief gradually giving way to low-angle accumulative abyssal-type topography. The lateral spread of these sediments suggests a system of submarine deep-sea cones tracing the base of the paleoslope.

The facies (landscape) assignment of packets in that case can be formulated as follows. Packets of the first type are the physical expression of facies of the principal (for a given complex) supply channels. With the distance from the base of the paleoslope, as a result of branching and attenuation of the dynamics, they give way to facies of secondary supply channels (packets of the second type). Given the initially low intensity of the transporting flows, the latter can also form independent channels. Packets of the third type, which occur in close association with the first two types of packets, are clearly interpreted as facies of wash-up bars bordering the supply channels. Their complex irregular structure and the frequency of landslide forms are easily explained by this interpretation (Fig. 3). Packets of the fourth and fifth types are the most common forms. They represent the true facies of the cone body and its most peripheral portions, but can also be associated with interchannel depressions. In the latter case, these packets are thin and alternate frequently with types 1-3 (Fig. 3).

We see from this description that packet types 1-5 were formed during the course of the sedimentary process of an autocyclic pattern (flysch packets). By contrast, packet types 6 and 7 do not belong to the autocyclic series. They suggest an irregular superposition of foreign factors, provoking a disruption of the rhythmic flysch accumulation (anomalous packets). Packet types 6 and 7 are identified by their different genetic character: they, and especially volcanogenic packets, are not connected with the transportation dynamic of the autokinetic flows that created types 1-5. The pulselike accumulation and peculiarity of occurrence bring these anomalous packets close to the rank of the primary genetic unit (a rhythm). On the other hand, the large space-time dimension, the heterogeneity, and the separateness of sediments are sufficient to assign to them the rank of packet. The classification of a particular packet as flysch or anomalous is important in analysis of the direction (cyclicity) of sediment accumulation. A complex section with flysch and anomalous packets (Fig. 3) can be represented as a histogram, with the thicknesses of packets plotted on the ordinate and their ranks on the abscissa. The ranks are expressed as the size of the sediment and characterize the dynamic of transporting flows. Let us compare the resulting histogram with an imaginary smoothing curve plotted without anomalous packets. In the former case, i.e., the actual section, an unstable dynamic is reflected, which disrupts bedding pattern as a result of landslides of unconsolidated sediments. On the histogram they appear as sharp irregular peaks. Remarkably, the highest density of these peaks is associated with facies of the principal supply channels. It can be attributed to the active hydrodynamic effect of downflowing streams. For anomalous packets associated with finer facies, tectonic factors are more likely. In the second case, the smoothing curve provides an assessment of the succession of facies settings at the level of fluctuations of the sedimentary process conditions under the definitive influence of migration of sediment-supplying channels. Generally, the introduction of anomalous packets does not affect the succession of facies. The packet themselves are not confined to any particular terrain.

The cyclicity (a directed succession of facies as a result of variation in the dynamic pattern of transporting flows) is represented in this case by two types: 1) a regressive cycle formed by the retreat of a channel with attenuation of the dynamics of transporting flows, followed by accumulation of thinner packets; 2) a progressive cycle, reflecting the advance of a stream with a gradually rising dynamic. The isolation of cycles by the presence of anomalous packets does not appear justified. This is also true of anomalous packets of volcanogenic origin (Fig. 4). Essentially, the cyclicity of true sediment accumulation is formed by general geologic factors (the Coriolis force, the equilibrium profile, etc.). Anomalous deviations against this background represent local catastrophic processes.

A relationship between these sediments and an island arc is beyond doubt. This is evidenced by the volcanic graywacke composition of the sediments with an andesite set of clastic minerals and the typical geochemical characteristics of the andesite-basalts, indicative of an island arc. The position of the sedimentary basin in relation to the island arc is much more problematic.

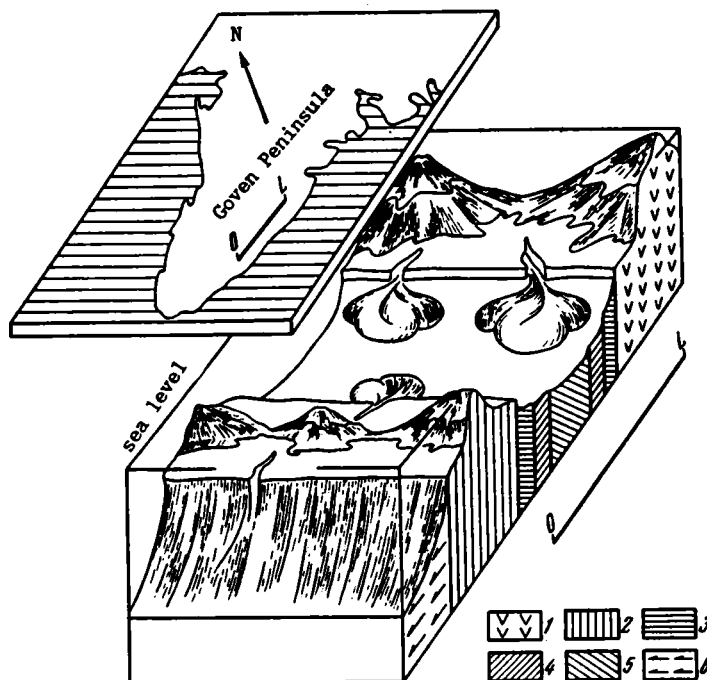


Fig. 5. A scheme of a forearc basin: 1, 2) alimentation regions (1 — island arc, 2 — volcanic rise); 3-6) facies (3 — marginal parts of the basin, 4 — central parts of fan systems, 5 — peripheral parts of fan system and central part of the basin, 6 — inner slope of the trench).

The following facts are important. Within the basin, the alimentation with clastic matter occurred almost exclusively from erosion products or the activity of the island arc itself. No influence of a continental supply source typical of rear basins is discernible. On the other hand, material with an oceanic association or low titanium shoshonites, which mark the rear portions of mature island arcs, are absent as well [3]. The fast variability of facies settings testifies to a strongly separated relief. The thicknesses and genetic types of sedimentary packets indicate fast-paced accumulation and the existence of intensive dynamic environments.

The position of the facies zones within the peninsula is remarkable. Two systems of submarine cones are determined. They are oriented subparallel to each other and trace the foundations of the slopes of two paleorises, expressed in the present relief by the Pylgin ridge (the northern system) and the Atavropel mountains (the southern system). Volcanic processes are well represented in the northern, more developed system. In the southern direction, the facies settings are gradually succeeded by forms reflecting a diminution of the intensity of transporting flows. The oblique and convolute lamination and orientations of landslide structures suggest that paleoflows and the general slope of the seafloor had a southern direction. In the southern system of the cones, a reverse picture is observed, but volcanic activity is completely absent. The southern system includes outcrops of ophiolitic rocks: a minor hyperbasite body uncovered in the core of one of the anticlines tilted westward whose wings are compounded by low-amplitude downthrows. The central part of the peninsula consists of facies of peripheral parts of the two systems.

The conditions which resulted in the formation of this set of characteristics in all likelihood developed in the frontal zone of an island arc. The presence of two subparallel systems and the exclusively island-arc composition of volcanites and sedimentary rocks and the deep-sea characteristics of the nanoplankton complex suggest that it is a forearc basin bounded by an active island arc on the northeast and a volcanic rise on the southwest (Fig. 5). A slight enrichment of the heavy fraction of Atavropel mountain rocks with sialic clastic minerals may be a consequence of dynamometamorphism, expressed during the course of formation of an accretional structure of a nonvolcanic rise. The widespread isoclinal folds indicate that the initial width of the sedimentary basin was at least three times its present width, attaining at least 150 km.

The formation of both valley-fan systems within the forearc basin started during a post-Maastrichtian period of volcanic lull as a result of destruction of the preexisting island arc of an ensiform structures. The main factors of sediment accumulation were denudation and autokinetic transport of clastic matter into the surroundings of the paleoslopes. The volcanic activation that occurred in the Upper Middle Eocene resulted in the northern system, in the superposition of products of submarine effusions upon volcanoclastic sediments of the submarine cones. In the southern system, the activation probably affected the geomorphology of the paleorise, intensifying denudation processes and stimulating further progradation of the valley-fan system.

This concept of the structure of the Goven forearc paleobasin is not inconsistent with the available data on the type and structure of basins near island arcs. As a modern analogue of the Paleogene Goven island arc system, one can cite the Marianas system: the structure of the active island arc, the forearc basin, and the nonvolcanic rise with ultrabasic rocks in the cores are in complete agreement with the concept of a Paleogene Goven forearc basin suggested by the analysis of these intensely deformed and fragmented strata.

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SUBCONTINENTAL FLYSCH IN THE CALEDONIDES OF THE MONGOLIAN-ALTAI-SAYAN REGION

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The composition, genetic types, and facies-paleogeographic conditions of the formation of Middle Cambrian—Lower Ordovician terrigenous flysch are studied. The flysch is classified on this basis as subcontinental. The specifics of this flysch type (flysch formation) are shown to reflect the characteristics of the alimentation provinces and the clastic matter supplied from them, as well as the morphology of the sedimentation basin. Ancient and recent analogues of the subcontinental flysch formation are cited.

A peculiar geosynclinal terrigenous formation is widespread in the Caledonides of Western Mongolia and the Altai-Sayan folded country. It includes the following series: mountain Altai series in the mountain and Mongolian Altai, and the Ustuishkin series in western Sayan and western Tuva. These thick (at least 5 km) terrigenous strata of a monotonic flyschoid structure cover immense territories, with a total area of at least 1 million km², and consists of uniform sialic clastics. They are present in the various structural-facies zones of the Mongolian-Altai-Sayan region [4-10, 12, 13, 16-18, etc.]. The flyschoid beds give way to faunistically characterized molassoids only in the northern and northeastern areas of the Altai-Sayan folded country [16].

The age of the terrigenous formation is established stratigraphically as the Middle Cambrian—Early Ordovician. In the western Sayan, it is succeeded with a gradual transition down the section by Vendian—Lower Cambrian terrigenous—silicic—volcanogenic rocks of the initial stage of Caledonide development and is covered by molassoids with Arenigian fauna. In the mountain Altai, the terrigenous formation occurs at the base of the Ordovician—Silurian strata. A similar situation is observed in western Tuva. In the Mongolian Altai it is the foundation on which younger superimposed structures developed in different tectonic blocks. The oldest sediments in some of them are Upper Ordovician—Silurian [10].

Thus, during the latter half of the Cambrian and the Early Ordovician an immense terrigenous sedimentation basin existed on the territory of the Mongolian-Altai-Sayan region. Its sediments exhibit all the lithologic characteristics of flysch [2] formulated by Vassoevich in the 1950-1960s. Original material was collected by the author in studies of these sediments in the Mongolian-Altai folded country; for the Altai-Sayan area data from the literature are used.

Faunistic Finds in the Terrigenous Formation

Both in the Altai-Sayan and the Mongolian-Altai folded country the formation is subdivided into two strata: the lower green unit and the upper variegated unit. These two strata are different in the composition and structure of the terrigenous rocks and in stratification patterns. The lower unit consists mainly of fine-grained sandstones, siltstones, and argillites. It is stratified unevenly. Members saturated with sand material display mainly sand—silt rhythms. Thin stratified members are formed mainly of silt-argillite rhythms. The upper unit is more coarse-grained, contains less pelitic material, and consists of larger sedimentary layers. It is also more variable in facies. The two units are different also in the composition of clastic matter. The specifics of the structure and stratification of terrigenous sediments in Western Mongolia have been described in detail in an earlier paper [3].

The most representative sediments of the lower unit are developed in the eastern part of the Mongolian-Altai ridge in the Kobdo zone, where the observable thickness varies from 1600 to 2300 m. In the northernmost part of this zone (near Lake Achit-Nur) at the center of the lower unit, one observes, besides the ordinary terrigenous rocks, thin layers (20-30 cm) of finely

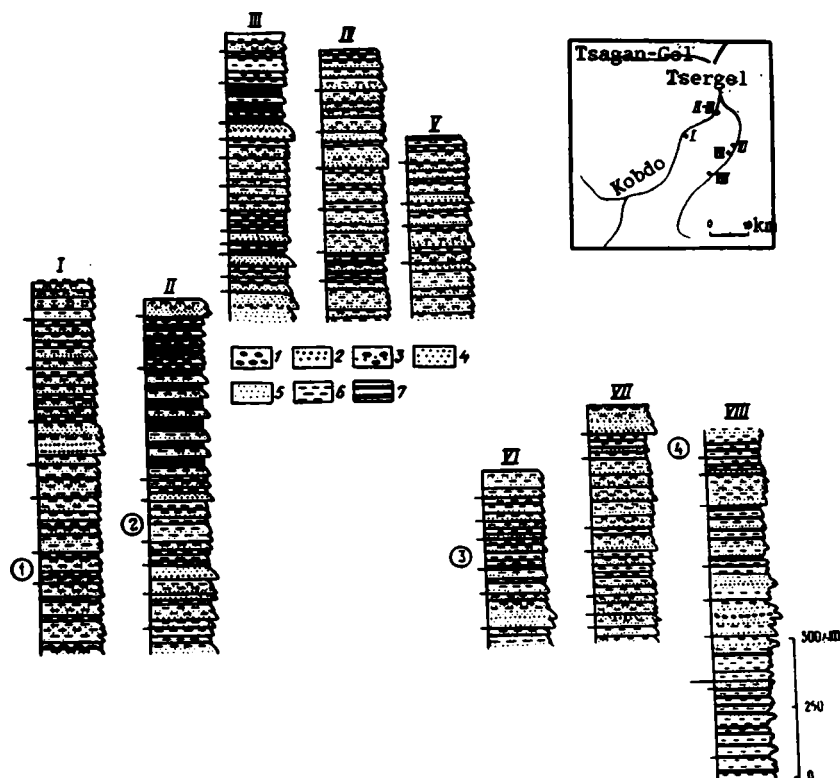


Fig. 1. Sections of terrigenous formation in the interfluvium of the Kobdo River and its right tributary, the Khargantu-Gol: 1) small pebble conglomerates; 2) gritstone; 3) gritstone with disseminated pebbles; 4, 5) sandstones (4 — coarse- and medium-grained, 5 — medium- and fine-grained); 6) siltstone; 7) silt, pelite, argillite. Circled numbers to the left of columns indicate location of vermiform burrows: 1) specimen 114/78; 2) specimen 60/78; 3) specimen 133/77; 4) specimens 123/77, 124/77; I-VIII sections and their location (inset).

clastic crystallovitric liparitic tuffs associating with clayey silicites and containing thin silicic sponge spicules, which rule out a Precambrian age of the enclosing rocks [15].

The rocks of the upper unit are most widespread in the Western Mongolian Altai in the Altai zone. In the interfluvium of the upper reaches of the Kobdo River and its right-hand tributary, the Khargantu-Gol, numerous vermiform burrows have been found in them. The main sites of collection are indicated in Fig. 1. In this area, the upper unit rocks are widespread, while the lower unit (observable thickness of at most 350 m) is exposed only at the core of the anticline on the left bank of the Khargantu River, 1 km upstream from the mouth of its right tributary, the Molon-Sala-Gol (bottom portion of section VIII). The vermiform burrows associated with the lower half of the upper unit belong to different genera (Table 1). They all represent the ichnofacies of *Nereites* (the group of burrows from the deepest-sea levels). An analysis of the burrows and their ecologic classification indicates that active colonization of the bathyal and abyssal regions began during the Early Ordovician [20].

Mineral Composition

Clastic rocks of the terrigenous formation belong to quartzose and micritic groups (in baricentric coordinates of a triangular diagram).

Most clastic formations belong to the quartzose group, whose rocks make up more than 50% of detrital quartz. It is followed by feldspars (mainly acid plagioclases) and lithic fragments. The quantitative proportions of the three components vary fairly regularly in plan. From north to south, the proportion of quartz grows while the abundance of rock fragments declines. In the heavy fraction, the presence of epidote decreases (Fig. 2). This change in rock composition is due mainly to the mineralogic disintegration of sedimentary matter during the course of sedimentation. A variation in the mineral composition is also observed from west to east. In the east (the Altai zone) lithic fragments are more diverse: microquartzites, quartzites, silicites, chloritized

TABLE 1. Forms of Traces of Bioturbation

Location (specimen number)	Taxon*	Age range
1 (114/78)	<i>Helminthopsis</i> cf. <i>tenuis</i> <i>ksiazkiewicz</i>	O-P
2 (60/78)	<i>Monomorphichnus</i> (?) <i>Palaeohelminthoida</i> sp. <i>Dictyodora</i> sp. <i>Crossopodia</i> sp.	C ₂ (?) Є - present C-C O ₁ -C
3 (133/77)	<i>Scolicia</i> sp.	Є - present
4 (123/77, 124/77)	<i>Protovirgularia dichotoma</i> M'Coy <i>Taphreimithopsis</i> sp. <i>Nereites</i> sp. <i>Crossopodia</i> sp. <i>Helminthoraphe</i> sp.	O-C O-C Є ₁ -P ₂ O-P ₂ O ₁ -C Lower range unknown

*Determinations performed by M. A. Fedonkin (Institute of Palaeontology, USSR Academy of Sciences).

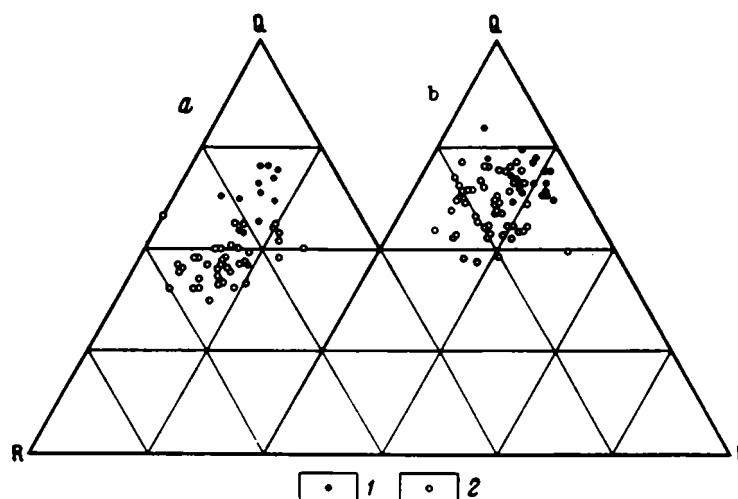


Fig. 2. Composition of medium- and fine-grained sandstones and coarse-grained siltstones of terrigenous formations in the basins of the Tsagan-Gol (a) and Khargantu-Gol (b) rivers: 1, 2) rocks from the lower and upper units, respectively.

fragments, occasional micaceous-silicic schists, argillites, andesites, ceratophirs, and aplites. In the west (the Kobdo zone), lithic fragments are less diverse: microquartzites and clayey-silicic and chloritized fragments. This is primarily due to the different petrographies of the alimentation provinces that supply the clastics to the different parts of the basin.

Mixtite rocks in the sediments of the upper unit contain less than 50% of each of the rock-forming components. In them, the proportion of clastic quartz also grows from north to south along the course of the zones, which is due to a decrease in the percentage content of lithic fragments. The composition of lithic fragments is fairly uniform on the entire territory. Mostly, sedimentary rocks are predominant: silicites (spongolites and jasper), clayey silicites, sandstones, and feldspathic-quartzose siltstones. In smaller amounts there are argillites, quartzites, and ferruginated, chloritized, and sericitized fragments. Ceratophirs, felsites, basalts, and plagiogranites are also found. From north to south, argillite and ferruginous and sericitized fragments disappear, and the amount of sandstones and siltstones decreases. Biotite appears in rocks of the mixtite group besides muscovite, which is characteristic for the quartzose group.

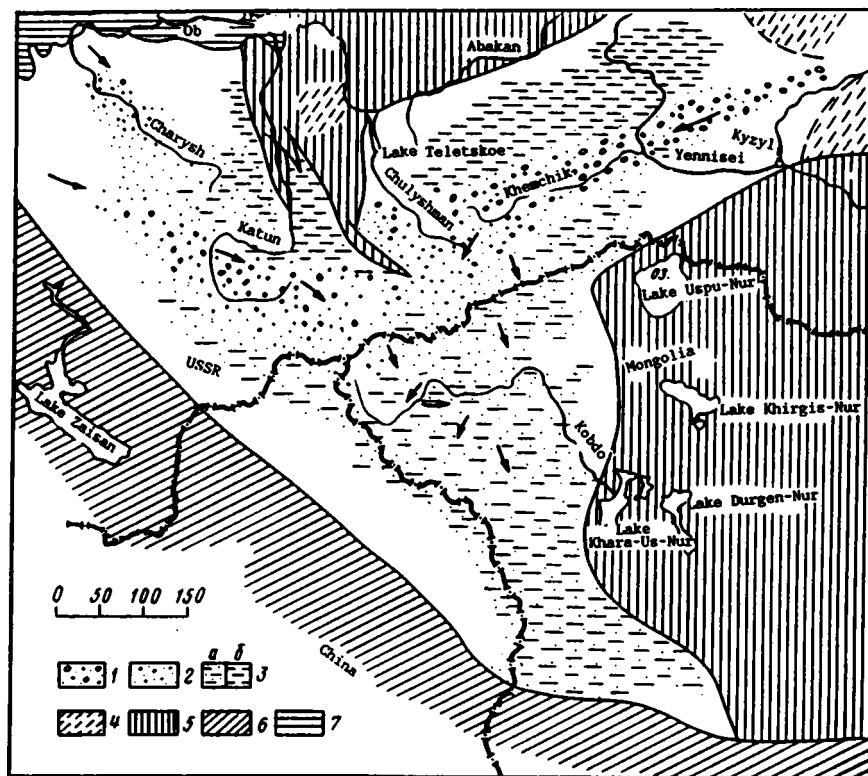


Fig. 3. Lithologic-facies schematic map for the end of the Cambrian/Early Ordovician Mongolian-Altai-Sayan basin (revised and updated from [3]): 1-3) bathyal sediments (1 — canyons (pebblestone, gravel, and sand), 2 — fans (sands of various structures, siltstone), 3 — inter-canyon spaces and plainlands (a — fine-grained silt clay, b — silt clay with sand interlayers)); 4) shallow-sea terrigenous sediments from large pebblestones to clays and oolitic limestones; 5) Early Caledonian and older folded surroundings; 6) Hercynian folded region (Irtysch-Zaisan and South Mongolia); 7) Cenozoic cover of the West Siberian Platform. Arrows indicate the direction of the transport of clastic matter (in the territory of Mongolia measured by base-layer proglyphs and obliquely stratified series).

The composition, stratification, and structure of the terrigenous formation of the Mongolian-Altai-Sayan region suggests that the principal sources of silicic material were outside the sedimentation basin, northwest and northeast of the current outcrops of these rocks. The main alimentation provinces were the Proterosayan landmass in the northeast and the Tobolsk* area in the northwest. They included terrigenous and silicic rocks, quartzites, silicic-clayey schists, granitoids (mainly plagiogranites) with subordinate felsites, andesites, and basalts. Other supply sources also existed, for example, in the present Katun-Abakan zone, where contemporaneous molassoids of a shallow-marine genesis were formed (Fig. 3).

Genetic Rock Types

Several genetic types are distinguished. They are mainly determined by the transportation and deposition mechanisms. Gravitites are most widespread, with a smaller presence of contourites and hemipelagites (Table 2).

Turbidites are predominant among the gravities. These are deposits of turbidity or suspension flows. Experiments and theoretical calculations indicate that they can transport silt and sand of various grain sizes. Gravels and pebbles are observed only as the "dragmat" of certain turbidites. Turbidites are characterized by a weak processing of material, which is unsorted or poorly sorted. This structural feature is more pronounced in coarser rocks. Basal cement is typical. According to Pettijohn et al. [14], they belong to the wacke group. Direct gradational distribution is the characteristic internal texture. The A, AE, AB, and

*Term coined from Sennikov [16].

TABLE 2. Main Features of the Genetic Types of Terrigenous Formation of the Mongolian-Altai-Sayan Region

Genetic type	Features of sediments							
	grain size of clastic material	amount of clay	sorting	bed texture	contacts	edaphogenic fragments	bed-thickness	mechanism retaining grains in a flow
Sediments of autokinetic flows (gravitites)								
Debrites	Pebble, gravel, sand	Large	Absent	Random, rarely slightly gradational	Both contacts sharp and uneven	Rare	0.2-1.5, less often up to 3.5	Forces of cohesion of clay particles
Turbidites	Sand from coarse to fine, silt; pebble and gravel in "dragmat"	Various (> 10%)	Poor, medium	Turbidite model of Bouma; direct sorting, fine horizontal, oblique, and convolute stratification	Sharp lower contacts with proglyphs with current jets, erosional with sculptures of loads, upper contacts even sharp and gradual	Wide-spread	0.05-0.9, less often 1.2 m	Inner energy of turbulent suspension
Sediments of hydrodynamic processes								
Contourites	Fine-grained sand and silt	Very little	Medium good	Massive, fine oblique low-angle and horizontal stratification	Both contacts sharp, lower often erosional	Both present	First millimeters, 7-9 m	Absent
Hemipelagites	Silt, pelite	Large	Ditto	Fine homogeneous	Lower contact often gradual, upper contact can be erosional	Absent	Diverse	

ABE intervals of Bouma turbidite model are frequent. Complete multilayers or multilayers with a reduced lower interval are observed less often. There is an abundance of surface textures (especially in the upper unit): mechanoglyphs and bioglyphs. Tongue proglyphs with a jet current of different dimensions and configurations are characteristic mechanoglyphs at the base of turbidite formations. Vermiform burrows are typical bioglyphs.*

The bulk of the gravel and pebble material was carried by inertial dense flows [19]. Coarse clastic layers belonging to the debrite group without stratification have been found among rocks of the upper unit. Near the base, the abundance of large fragments is increased. Sharp and uneven lower and upper contacts are typical. The material has been worked to different degrees. It is unsorted, and sometimes subparallel orientation of the clastic units is observed. The binding groundmass contains a considerable proportion of clay matter.

The hydrodynamic processes formed contourites, represented by siltstones and medium- to fine-grained sandstones in layers of different thickness but not thicker than 10 cm. They contain very little clay (film and pore cement). Sorting is medium, sometimes good. Fine horizontal and oblique stratification and lenslike beds are typical. The lower and upper contacts are sharp.

Fairly large interlayers and members of argillites and silt-pelites (a mixture of clay and silt in approximately equal proportions) are found amid gravitites, interpreted as hemipelagites. The rocks are usually schistose and cleaved, and have internal texture. It is impossible to determine whether they were deposited from a nepheloid layer or from a water body, i.e., whether they are clayey contourites or decantites.

*See [2] for details on the structure and texture of terrigenous accumulations and turbidite beds.

Generally, bathyal sedimentation conditions are postulated for this formation according to the following criteria: first, vermiform burrows of *Nereites ichofacies* (the deepest group of traces); second, a leading role of turbid sedimentation (no textures of shallow zones observed); third, carbonate-free formation, determined either by the bathymetry of the sedimentation basin or the climatic factor. The climatic conditions during the formation time of terrigenous strata did not rule out deposition of carbonate, as evidenced by contemporaneous shallow-sea facies developed on the northern periphery of the basin (in the Altai-Sayan region) and containing lenses of oolitic limestones and calcareous organism remains. The absence of sedimentary carbonate in the rocks of this formation therefore is due to the fact that they developed below the paleolevel of carbonate compensation, which is the Paleozoic lay at lower depths than in the present oceans [1].

The Mongolian-Altai-Sayan paleobasin had a complex configuration but was generally oriented submeridionally (see Fig. 3). Its true size cannot be established at present; however, considering the areas of current terrigenous formation outcrops, it was longer than 1000 km with a width varying from 300-500 to 800-900 km. The author's data on the Mongolian-Altai folded country and literary data on the Altai-Sayan region indicate that deep-sea terrigenous sedimentation was predominant on this large territory. The composition of clastic material, the genetic type of the sediments, the specifics of their stratification, and their distribution in plan suggest that these rocks were formed in subaqueous valley-fan systems on intercanion areas and bathyal planes.

The thickest and longest valley-fan system existed in the northeastern part of the basin (the upper reaches of the Bashkaus River and the valley of the Khemchik River). It was oriented in a northeasterly direction. Generally, from southeast to northeast the amount, thickness, and grain size of clastic matter increase in this strip. The material enriches portions of the sections, which are coarse and consist of large layers. They are separated by finer-grained varieties with thinner stratification. This uneven deposition is evidence of the complex evolutionary history of the valley-fan system, largely determined by the initial clastic material supplied from the alimentation province. Fine clastic and clayey sediments with thin and fine interstratification were predominant on intercanion areas.

The distal portion of this valley-fan system is traceable in the northeastern part of Western Mongolia (the Khub-Usu-Gol valley). The lower bed is formed of distal turbidites corresponding to sediments of the basin plane and the outer fan. The basin sediments are represented by a thin interstratification of horizontally laminated or layered beds of fine-grained sandstones and siltstones (1-5 cm) with silt pelites (3-10 cm). Gradationally sorted sand-silt interlayers are also found. Outer fan facies are prevalent in the section. These are rhythmic alternations of medium- to fine-grained sandstones, siltstones, and silt pelites. The thickness of the rhythms varies from a few centimeters to 35 cm, sometimes up to 0.6-0.7 m, and even 1.2 m in individual cases.

Turbidites typical of the higher portions of alluvial cones are mainly found in the upper unit of the terrigenous formation. River-bed facies (subaqueous channel formations) are most common. They are represented by rhythmic interstratifications of layers with gradations from coarse- and more often medium-grained sandstones to siltstones, and rarely silt pelite. They vary in thickness widely from 10-15 to 90-120 cm. The typical sequences are A, AE, AB, and ABE of the turbidite model.

In the northwestern part of the basin, valley-fan systems oriented from northwest to southeast also existed. The sediments of the deep-sea alluvial cone in one of them are traced in the Western Mongolia Altai, in which the direction of the transport of clastic matter by autokinetic flows varied substantially. This was probably a result of meandering subaqueous riverbeds of the valley-fan system. The existence of two principal sources of clastic material is also confirmed by the differences in the mineral composition of clastic matter in the western and eastern parts of the paleobasin. Southward, the alluvial cone rocks are succeeded by sediments of the bathyal plane formed of fine and small clastic and clayey material.

Generally, the terrigenous formation displays a tendency for the clastic matter to increase in grain size and to acquire more red pigmentation from the bottom to the top of the section: the upper unit is coarser, with predominant polymictic clastics and a larger amount of iron oxide than in the lower unit. The general vertical orientation of lithologic variability within the formation presumably reflects a change of landscape in the alimentation provinces.

The flysch formation of the Mongolian-Altai-Sayan region was formed as a result of activation of tectonic motions in the continent-paleobasin system. Major tectonic events, established at least in the northeastern margin of the paleobasin, occurred during the Middle-Early Late Cambrian [11, etc.].

In the Eastern Sayan and areas further east, they created a large continental mass with numerous granitoid batholiths. Similar events probably took place in the northwest margin of the flysch basin. The erosion of these continental masses was mainly responsible for the creation of the terrigenous formation, which can be described as subcontinental flysch.

Accumulations of a similar composition and stratification are found in southeastern Canada, Nova Scotia, known as the Meguma group. These are thick rocks (~10 km) represented by quartzose graywackes, metawackes, and schists. They are dated at the Middle Cambrian—Early Ordovician. Two strata are distinguished in the Meguma group. The lower bed (sandy flysch) was formed by an interstratification of medium- and small-grained sandstones and finer rocks. The rhythms range in thickness from 1 to 50 cm, and the layers are uniform or gradationally sorted, often horizontally layered. At the base, some of the rhythms contain gravel-pebble material. Heiroglyphs, both mechanical and biogenic, are typical. Landslide textures are abundant. The top bed is of a finer-grained material (with predominance of schistose flysch facies). It consists of siltstones and schists with sandstone interlayers. Beds with CDE and BCDE intervals of the Bouma turbidite model are prevalent, while mechanoglyphs are less common [21, 23, 24].

The Meguma group occurs at the base of a geosynclinal section and is covered conformably by the White Rock series. Most of the formation evolved in a deep-sea setting, but its uppermost portion was formed in a relatively shallow-sea environment. The sandy accumulations of the lower portion belong to sediments of a deep-sea canyon. A considerable part of the upper bed consists of sediments of an abyssal plane. The uppermost layers of the Meguma group and the covering rocks were formed within a continental shelf. Sialic clastics were brought into the deep-sea basin from the outer Morocco shelf. The Meguma group was a segment of the Mauritanian eugeosyncline, located between the coastal Meseta Morocco and the Spanish Sahara-Mauritania.

In contrast to the terrigenous formation of the Mongolian-Altai-Sayan region, with its general trend for clastic material to be coarser with time, Meguma rocks display an inverse succession in the alteration of the granulometry of clastic material. It reflects the evolutionary patterns of landscapes of the alimentary provinces, primarily different times of activation of tectonic motion in the continent—paleobasin system.

Atlantic continental margins are modern analogues of the subcontinental flysch formation. Canyon and valley-fan sediments at the base of the continental slope in the western part of the North Atlantic and the related Som and Hatteras abyssal planes, briefly described in [2], are especially informative.

In the northern part of the Indian Ocean, the sediments of the Bengal deep-sea alluvial cone belong to this group of formations. Borehole 218 was drilled in that cone 18 km west of the principal fan valley [26]. The total thickness of sediments penetrated is 773 m. They cover an age interval from the Middle Miocene to Recent. Quartz and feldspars are predominant in clastic matter (up to 75%), which also includes sediments of rocks, mica, and accessory minerals (hornblende, epidote, garnet, pyroxene, tourmaline, and metalliferous units). The bed consists of greenish-gray and gray sandstone, siltstone, and their sandy and clayey varieties, as well as nanofossil ooze.

The bed is unevenly stratified. At the top of the section (from 0 to 350 m), sand and silt beds are thickest (up to 50 cm) and display direct gradational sorting. Clastic material is angular and semirounded. The upper and lower contacts are sharp. At the center of the section (from 350 to 500 m), the sandy beds become thinner (to 5 cm). Fine well-sorted silt interlayers (up to 1 cm and less) and nanofossil chalk with bioturbation sometimes occur. The bottom of the section (500-773 m) is built rhythmically: an alternation of gradationally sorted sandy siltstone (<15 cm), horizontally and obliquely stratified siltstone (up to 2 cm), and clayey siltstone.

Four types of turbid activity are distinguished according to the distribution of the dominant grain sizes in terrigenous clastics and pelagic material. They are associated with the Middle/Late Miocene and the Early Pliocene and Pleistocene. The intensification of turbid sedimentation can be caused by three factors: 1) migration of subaqueous distribution channels over time; 2) climatic variations resulting in a change in sea level and, in turn, affecting the rate of influx of clastic matter into the submarine valley-fan system; 3) tectonic activity in the Himalayas, which intensified erosion in the Ganges—Brahmaputra River system, that supplied clastics to the Bengal deep-sea alluvial cone. The Pleistocene activation of turbid sedimentation was caused by a glaciation maximum which lowered sea level and by the Late Pleistocene and Pleistocene—orogenic motions in the Himalayas. The Late Miocene orogeny in the Himalayas also intensified turbid sedimentation. The oldest stages of its activation were caused by migration of the subaqueous distributive channel in the deep-sea alluvial fan.

The silt—sand and sand—clay flysch is typical for sediments of abyssal planes and interchannel facies of alluvial cones. In subaqueous canyons and channels the grain size categories of terrigenous clastics and the thickness of turbidite beds are in direct correlation with the initial clastic material characteristics (grain size and quantity). Besides sand, coarser varieties may be present in valleys. In many canyons (Hudson, Monterey, La Jolla, Redondo, etc.) gravel and pebbles have been found. In Zambesi Canyon (borehole 243), at a depth of approximately 4 km, 800 km from the estuary of the Zambesi, coarse sand and gravel were

found (measuring 1-2 mm to 1.5 cm across). The material is semirounded or well-rounded and consists of quartz (95%), feldspar (2%), and rock fragments (granites and quartzites) [25]. Edaphogenic boulders which have fallen from the walls of slope valleys are sometimes found in the canyons.

At the transition of abyssal valley facies and distal portions of deep-sea fans to the facies of their proximal areas and then to river course sediments of distributive channels and canyons, the clastic material becomes coarser, the turbidity units grow in thickness, and the overall stratification pattern of terrigenous accumulations is changed. Thus, a lateral transition is observed from finely stratified clayey flysch to increasingly coarser clastic varieties, up to gravel-pebble materials with horizons of a landslide-collapse genesis. The granulometric and quantitative changes in the clastics over time, the meandering of submarine valleys, their progradation, and other processes, produce unevenly stratified flysch beds of a complex structure.

Specifics of Subcontinental Flysch

A considerable body of factual data obtained in the past decades both on recent oceans and ancient geosynclinal systems indicates a predominance of deep-sea rocks in these structures. The principal genetic types and facies belong to gravities, mainly turbidites, which create rhythmically layered flyschoid strata. These sediments are formed in various geodynamic environments during different stages in the development of geosynclinal zones. Accordingly, the concept of flysch should be limited and applied only as a synonym for flysch formation, confined to a specific stage in the development of geological structures.

Besides the classical flysch formed in trough basins and associated with intrageosynclinal orogenesis, subcontinental flysch (subcontinental flysch formation) should be identified. It was formed in extensive morphologically complex basins mainly under the effect of extrageosynclinal orogenesis [22], i.e., major tectonic events in the continent-sea basin system. Its features include: sialic composition of detrital clastics supplied by the continent, formation at the base of a continental slope and in adjacent deep-sea marine basins, with transport of clastic matter mainly through valley-fan systems, and development of deep-sea alluvial cones and basin sediments. A "small-scale" facies variation of sediments reflects, primarily in the granulometry of the clastic matter and the textural characteristics of sediments, the widespread occurrence in plan and large thickness of units determined by avalanche turbid sedimentation.

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DEPOSITS OF DENSITY CURRENTS IN THE SETTE-DABANA CARBONATE COMPLEX (SOUTHERN VERKHOYAN'E)

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UDC 551.3.051:552.54:551.73(572.56)

We identified deposits of avalanches and slides as well as debris, liquified, and turbidity flows in carbonate strata of Middle—Upper Cambrian and Lower Carboniferous. Evidence of bottom currents was found. Sedimentation took place in a sloped-shelf environment during the Middle—Late Cambrian and on submarine fans during the Early Carboniferous. In both cases, the depth of the basin exceeded 1.5 km.

Numerous investigations of depositional environments since F. Kunen shown the important role of density (primarily turbidity) currents in the accumulation of recent and ancient clastic strata. Density currents of a carbonate composition are widely occurring at the margins of the present-day continents, but they were once considered to have had only limited distribution in ancient basins of carbonate deposition (with the exception of the Mesozoic Tethys basin) [12, 13] In the last 10-12 years, however, deposits of different types of density currents were found in pre-Mesozoic carbonates of Thallaskii Alatau, the Urals, the Cordilleras, and other regions of the world, including Sette-Daban. According to the findings of the authors, density currents of carbonate composition are recognizable here in sediments from the Middle Cambrian to the Lower Ordovician and in the Lower Carboniferous. The sediments of the Cambrian and Carboniferous are of the most interest, since their widespread distribution (Fig. 1) allows us to infer structural features of the depositional basin over a significant area.

The region discussed in this report is located at the southeastern margin of the Siberian platform, has experienced several periods of tectonic activation, and now has a complex fold-overthrust structure. To reconstruct the original layout of sections brought together by folding, the authors used the equal-area method of balancing sections, one of the most widely used methods in modern structural geology [18]. We used the criteria and terminology suggests by the American Association of Petroleum Geologists [15] to characterize mass-transport processes. They distinguish avalanches, slides, and sediment gravity flows; the last two are broken down into more detailed subclassifications. We used deposits of density currents as a more general term, since the term gravitite, which is often used in our scientific literature [8], has a more particular meaning in our classification. The classification of R. Dankheim was used for microscopic descriptions of carbonate rock, and the classification of J. L. Wilson was used for standard microfacies [12]. A capital letter T with a letter subscript indicates a corresponding division of a Bouma sequence (T_a , T_b , ...) or its structure (T_{abcd} , T_{ac} , etc.). Five divisions (T_a , ..., T_e) were initially recognized in the Bouma sequence. Subsequent investigators added a sixth division: T_f , representing pelagic or hemipelagic sediments. Capital letters of the Latin alphabet (A, B, ..., G) denote facial divisions according to Mutti and Ricci-Lucchi lying at the base of one of the most widely used sedimentological models (Fig. 2).

Deposits of Middle—Upper Cambrian Density Currents

A diagram showing the stratigraphic dismemberment of the Middle—Upper Cambrian sediments is shown in Fig. 3. Below, we will devote primary attention to the Ulakhsk series and the eastern area of outcrops belonging to the Ust'-Maisk and Kerbinsk suites, where up to 70% of the entire section has a flysoid structure and is represented by rhythmically alternating carbonates and argillites.

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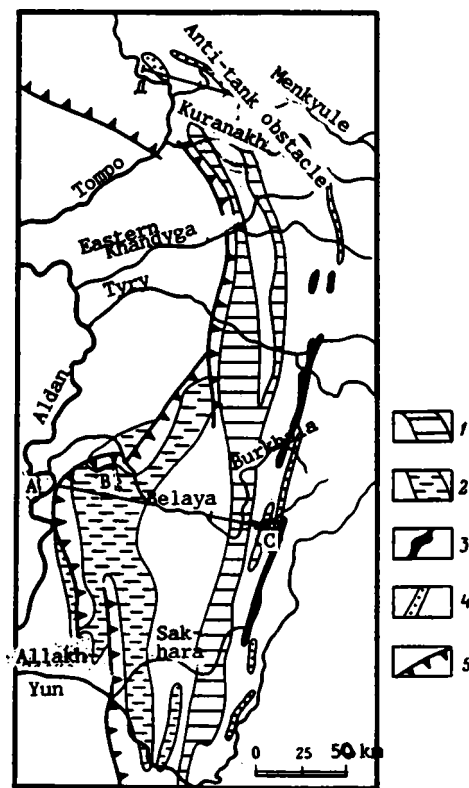


Fig. 1. Map of outcrops of Middle and Upper Cambrian and Lower Carboniferous. 1-2) outcrops of sediments of the Middle and Upper Cambrian (1: Ulakhsk series, 2: Chaisk, Ust'Maisk, and Kerbinsk suites); 3-4) outcrops of Lower Carboniferous (upper Tournasian-Middle Visean) sediments (3: Kur-anakhsk suite, 4: Muruninsk and Menkyulensk suites and their analogs); 5) Nel'ano-Kyllakhsk junction separating the Siberian platform from the fold-overthrust structures of Sette-Daban; A, B, C) locations of paleogeographic profiles.

Four divisions of an entire Bouma sequence can be distinguished on the basis of textural features in the most complete sequences. In the lowest division (T_a), the carbonates are homogeneous, sometimes with indistinctly graded bedding. Upward, the rocks are horizontally bedded and graded bedding is rare (T_b). The third division (T_c) is characterized by gentle cross-lamination and indistinct wavy lamination. Such complete sequences are rare and apparently characterize only the Bilyakhsk suite of the Eastern Khandyga river basin. Sequences of types T_{ae} , T_{abc} , and T_{ace} occur more often. Division T_e is occasionally absent. The thickness of a sequence in the Ulakhsk series is usually 15-20 cm, but varies from 3-5 cm in the Elovsk suite to 1-1.2 m in the Bilyakhsk suite. In the Ust'-Maisk and Kerbinsk suites, the thickness of a sequence rarely reaches 15-20 cm; most often, it is 3-7 cm. Thin sequences usually exhibit the structure of T_{ac} . The limestone/argillite ratio in the Ust'-Maisk and Kerbinsk suites varies from 1:3 to 3:1; this ratio almost always exceeds 3:1 in the Tisovsk and Bilyakhsk suites, and often equals 5:1-7:1. The contacts between sequences are usually smooth; occasionally small load casts and imprints of nodules are noticeable at the bottoms. At the tops of sequences lacking T_e , there are sometimes wave ripple marks; this is characteristic of certain types of carbonate deposits of gravity flows [15].

A significant portion of the limestones were subjected to dolomitization, quartzification, and dissolution under the influence of recrystallization, and were converted to inequigranular rock without relicts of primary structure as a result of recrystallization. In a number of cases, it was found that the limestones have a primarily fragmental texture and are easily distinguishable from the groundmass on the basis of darker color. The rock clasts are primarily rounded, and their size varies over a wide range, reaching 0.6 mm. Prolate clasts are usually more poorly rounded and show uniform orientation; more isometric grains are chaotically oriented. The sorting of clastic grains is usually poor, but improves somewhat in the upper

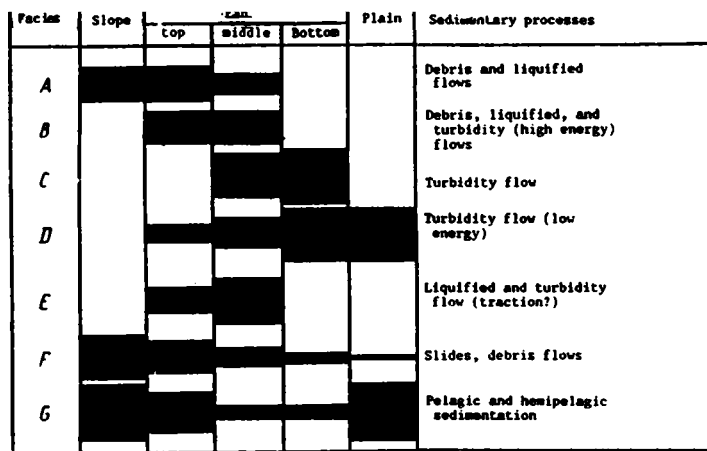


Fig. 2. Distribution of facies of Mutti and Ricci-Lucchi in ancient submarine fans. Thickness of bar corresponds to frequency of occurrence of a given facies.

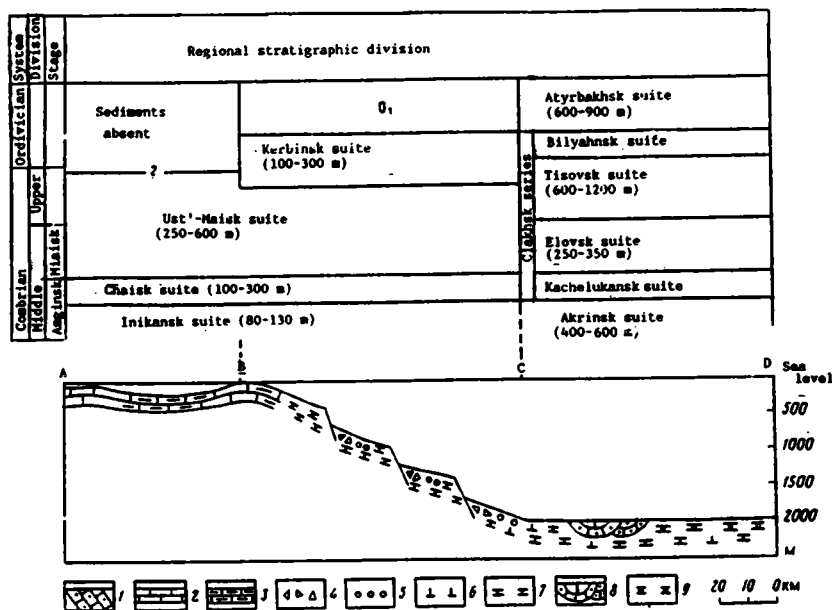


Fig. 3. Diagram of stratigraphic dismemberment of Middle and Upper Cambrian sediments (compiled according to [2, 10, 14]) and paleogeographical profile for central portion of Lower Cambrian. Profile compiled from palinspastic reconstructions; placement of letters on profile corresponds to placement of letters in Fig. 1. 1) alluvial-deltaic cross-laminated calcarenites with variegated coloration; 2) shelf limestones; 3) clayey limestones; 4-7) deposits of density currents on slopes and feet; 4) slides; 5) debris flows; 6) liquified rocks; 7) predominantly carbonate turbidity flows; 8) deposits of bottom currents (cross-laminated calcarenites); 9) pelagic limestones of oceanic basins.

portions of the sequences. The clastic grains have incrustate rims cementing a mass of calcite completely recrystallized and converted to inequigranular crystals, with small unevenly distributed rhombohedrons of dolomite. Packstones predominate, based upon the ratio between grains and groundmass; wackestones are more rarely identifiable. Laminates, alternating layers of limestone and argillite forming sequences from a few millimeters to 3-5 mm in thickness, are present at outcrops of the Ulakhsk series along the Belaya River (below the mouth of the Burkhala River), both within Bouma sequences and in the form of packets up to several meters thick. Limestones occur in the lower portion of a sequence and are represented by fine inequigran-

ular greenstones-packstones with calcite of 0.05-0.15 mm grain size. Up to 10% of the rock is composed of clastic quartz and silt-sized feldspar. Upward along a sequence, the quantity of carbonate material decreases rapidly but uniformly; clastic quartz disappears, and the role of clayey material increases. This creates similarities to graded bedding. There are no recognizable textures within a sequence. The lower contacts of the limestone layers are sharp, uneven, and with traces of intrusion of calcareous material into unlithified clayey sediment.

Beds of breccias and breccia-conglomerate are widely occurring in the Kerbinsk suite in the middle courses of the Belaya and Chukhonoi rivers. Clasts are primarily elongate and sharply angular. There are fragments of the layers up to 80 cm long and less than 10 cm thick. Poorly rounded and rounded fragments of light marls and dark-gray limestones absent in the enclosing rocks can be identified. Very small clasts are chaotically oriented; larger clasts are oriented approximately parallel to one another. Clasts freely "float" in the clayey—calcareous mass surrounding them, a mass similar to the clasts in composition and clearly distinguishable only by weathered surface. One can find textures similar to crushed submarine slide sets. The thickness of breccias varies from several centimeters to 15-20 m. They have lenticular shapes, and thin (up to 0.5-0.8 m) bodies pinch out over a distance of several meters. Contacts of the breccia bodies can differ, being either gradual or sharp. In the latter case, the enclosing rocks around the contacts are sometimes compressed into small folds the quantity of breccia decreases in an eastern direction; breccias compose comparatively thin bodies at the western outcrops, and are generally absent further to the east. The quantities of breccias in the lower reaches of the Belaya River and at exposures along the Aldan River decrease significantly.

In sections of the Ulakhsk series, there are carbonate rocks of still two other types in subordinate quantities. The first type are clayey—calcareous mudstones and wackestones with calcite crystals no more than 0.03 mm in size. The irregular distribution of clayey material gives the rock a lumpy appearance. Comparatively pure carbonate is concentrated in crescentiform or circular shapes up to 0.2 mm in size, possibly recrystallized remains of microfauna and fine detrites of thin-walled shells (standard microfacies 3). The second of the types of rock mentioned are calcarenites widely occurring in the Tisovsk suite in the Eastern Khandyga river basin. In comparison with the calcarenites from the lower portion of the sequences, they are noticeably rich in clastic quartz and are poorly sorted. The fragmental grains do not exceed 0.2-0.3 mm in size; the cementing clay mass constitutes no more than 10-15%. There are packets of cross-laminated rock from 5-7 cm to several meters in thickness. Plane-parallel cross-lamination predominates; the thickness of an individual series does not exceed 15-20 cm. Serially arranged junctures are distinctive, subhorizontal or slightly inclined in the direction opposite the dip of the cross laminations themselves. Such cross-lamination forms during a hydrodynamically active regime creating straight or slightly curving beds [6].

A schematic paleogeographic profile for the Late Cambrian is shown in Fig. 3. In the western area, there are exposures of rhythmically interbedded limestones, marls, and argillites with diverse fauna and massive limestones with algal buildups [2] this is characteristic of the open shelf and lagoons with free exchange between different waters. At the shelf margin, according to the data of V. A. Yan-zhan-shin [14], who found desiccation cracks and traces of raindrops (clear signs of periodic drying of the portion of the region under consideration), there is a rise. East of the rise, a slope begins toward a deep-water trough. The clearly lentiform shapes of the breccia bodies and the abundance of intraformational surfaces among thin T_{ac} rhythmites cross-cutting lower lying sediments at angles of 15-20° suggest the presence of numerous submarine channels here. The breccias described above are characteristic of deposits of slides or the very densest of gravity flows, debris flows [15]. Breccias and calcareous—clayey (rhythmites characterize facies F and G in the model of Mutti and Ricci-Lucchi. This association occurs only on slopes, especially on their upper portions [15, 19]. Toward the east (the region of the Ulakhsk series, see Figs. 1, 3) there are no deposits formed due to slides of material, and the role of debris flows is sharply reduced. Deposits of liquified flows are identifiable only in rare cases on the basis of convolute lamination or characteristic intrusion textures in unlithified sediment. The major portion of the section of the Ulakhsk series is composed of turbidites with reduced Bouma sequences. These sediments characterize facies A, C, and D. If we take into account the presence of packets of laminates (facies G), the set is typical of the middle portion of a fan [19]; in the absence of facies J (debris and liquified flows) the set is typical of the lower portion of a fan [19]. Micro-lumpy pelagic limestones most likely accumulated on the periphery of fans or on the abyssal plain. The boundary between outcrops of the Chaisk, Ust'-Maisk, and Kerbinsk suites, on the one hand, and the Ulakhsk series, on the other, is characterized by changes in thickness and composition of sediments. West of this boundary, slide movements of masses with elastic or plastic properties played an important role; east of the boundary, liquified and turbidity flows with properties similar to the properties of a liquid played important roles [15]. This boundary apparently corresponds to an ancient boundary between slope and foot at which abrupt impact of density currents occurred. The earlier-mentioned cross-laminated calcarenites of the Tisovsk suite, judging from the dimensions of the crosslamination, accumulated under the influence of northern or northeastern currents approximately parallel to the boundary of the slope and its foot. Paleocurrents of similar direction were found south of the region under discussion in the Mai river basin [11]. According to the criteria proposed in [17, 21], these

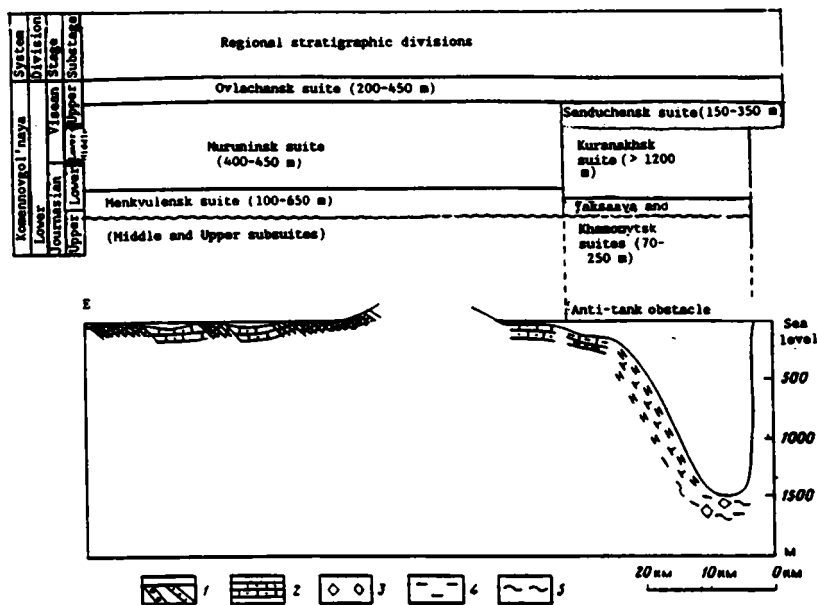


Fig. 4. Diagram of the stratigraphic dismemberment of the Lower Carboniferous deposits (compiled on the basis of [4, 5, 14]) and a paleogeographical profile for the second half of the Tournasian. The profile was composed on the basis of palinspastic reconstructions; locations of letters on the profile correspond to locations of letters in Fig. 1: 1) alluvial-deltaic conglomerates; 2-3) shelf deposits (2: interbedding of calcarenites and limestones, 3: clayey limestones, rhythmic interbedding of limestones and clayey limestones); 3-4) deposits of density currents on slope and foot (3: avalanches, 4: predominantly clayey turbidity flows); 5) phthanites (deposits of deep-water troughs). For remaining designations, see Fig. 3.

calcarenites show a similarity to present-day silty-sandy deposits of bottom currents (contourites) and probably formed as a result of erosion of deposits from density currents.

Transport of clastic material took place along submarine channels characteristic of submarine fans. However, the lentiform breccia bodies filling the submarine channels are united into packets traceable along the strike of the structure for no less than 200 km; this suggests a linear shape for the source of drift. This similarity of depositional environment to both fans and slope apron allows us to classify it as a hybrid type, in the terminology of B. Stow et al [22]. Apparently, present-day analogs to this type of environment are known only for terrigenous turbidites.

The trough (Fig. 3) has a submeridional strike, and the width of its visible portion amounted to 70-80 km. Based upon palinspastic reconstructions, the width of the slope is no less than 70-90 km. Since movement of slides and debris flows usually begin around steep scarps and continue along a slope with angles of 1° or less [15], the depth of the basin might be approximated as 1.5 km. The thickness of the complex filling the deep-water trough (from the deepest-water pelagic shales to the nearest stratum of shallow-water limestones) equals 1.5 km, but, if correction is made for rock compaction during lithification and as a result of deformation, the depth was approximately 2-2.5 km [16, 18]. It is precisely these values which determine the probable depth of the sedimentary basin.

Deposits of Density Currents of the Lower Carboniferous

A diagram showing the stratigraphic dismemberment of the deposits of the Lower Carboniferous is shown in Fig. 4. Below, we will devote our primary attention to the Kuranakhsk suite, with rhythmically alternating limestones and argillites of flysoid appearance predominating in its section.

On the basis of textural criteria, all of the divisions in the Bouma sequence can be distinguished in the most complete sequences: gravationally bedded (T_a), with thin horizontal bedding (T_b), gentle cross-bedding and wavy bedding (T_c), with fine horizontal, often graded bedding (T_d), and uniform unbedded (T_e). Divisions T_a - T_d are composed of sandy and silty carbonates, while T_e is composed of argillites. Divisions T_a and T_c are the most clearly expressed; sequences of the type T_{acc} or T_{acdc} are the

most frequently occurring. The thickness of a sequence varies from 2-3 m to no less than 1.5-2 m, with a predominant value of 15-80 cm. As a rule, limestones predominate over argillites (2:1 and more), but thick (no less than 80-100 m) packets in which the limestone/argillite ratio does not exceed 1:10 are present in all of the sections. The contacts of the sequences are uneven; load casts, imprints of nodules, and recesses of elongate triangular shape are sometimes noticeable at the bottoms.

The deposits of the Carboniferous are generally less altered than those of the Cambrian; one can recognize primary structures and textures, although secondary processes are manifest here, including dolomitization, quartzification, and albitization. Within sequences, the majority of the carbonates have a fragmental texture and are lithoclastic and biolithoclastic calcarenites. Lithoclasts consist of clayey limestones and are usually well rounded. Lithoclasts vary in size from 0.05 to 0.3 mm. Bioclasts are represented by fragments of crinoids and, more rarely, microfauna, various shells, and algae. Bioclasts reach 1 mm in size. In comparison with the lithoclasts, they are less well rounded and almost unsorted. The quantity of bioclastic material varies from a complete lack to 60% of all clastic material. Clear incrustate rims and traces of later dissolution are present. Clayey—calcareous cement is recognizable in the least altered samples. In terms of the ratio of fragmental grains to groundmass enclosing the grains, wackestones predominate. The sorting is poor, but improves somewhat toward the top of a sequence. Fine laminations in divisions T_b , T_c , and T_d are due to concentrations of clayey material in layers no thicker than 0.5 mm. In general, thin horizontal lamination is not as well expressed as cross-lamination. Calcarenites contain significant quantities of quartz clasts, sparse clasts of feldspars, and individual clasts of sphene and zircon. The quartz clasts are of the same size as the carbonate clasts, but are not as well rounded; they usually make up approximately 30-40% of the entire rock in division T_d . A portion of the fine laminations have a clearly graded structure as a result of changes in content of quartz clasts and clayey material.

The final sequences of argillites (division T_e) usually contain carbonate and siliceous material. Quartz fragments are absent or make up a few percent of the rock. The rocks are homogeneous and unbedded, but sometimes an uneven distribution of carbonate material gives them a blobby-lumpy appearance.

Dark-colored clayey—siliceous wackestones with numerous uniformly oriented sponge spicules and sparse radiolarians are widely occurring in packets of essentially clayey composition. These rocks are pelagic (division T_f) and belong to standard microfacies 1 (a basin facies [12, 15]). It should be noted that the boundary between divisions T_e and T_f is almost undeterminable under field conditions. As the amount of siliceous material decreases, the carbonate content and the wackestones grade into the phthanites described in detail by M. D. Bulgakov [1]. In the Kuranakhsk suite, there are silicites formed from tuffo-argillites and also tuffs of andesites and liparites in addition to phthanites [1]. They are characterized by light color and compose beds up to several tens of meters thick which are well maintained along strike. A portion of the silicites formed from silicification of limestones; this can be deduced both from flint nodules widely occurring in the limestones and from the replacement of gradationally laminated calcarenites with silica up to the point of conversion to siliceous rock with relicts of graded lamination (observable in thin sections).

Calcarenites are present not only in Bouma sequences, but also in the form of separate beds ranging from several tens of centimeters to 2 m in thickness. Clasts vary from silt-sized to gravel-sized; a high (30-40%) content of clastic quartz is present. The most typical structure has convolute bedding. Submarine-slide folds and numerous flame-type structures resulting from intrusion of quartz—carbonate detritus into unlithified, essentially clayey sediment are associated with these layers. Packets containing numerous beds of calcarenites are well maintained along strike, but the beds themselves are apparently lentiform.

Coarsely fragmental rocks occur in minor quantities; they are found at the periphery of outcrops of flyschoid strata. According to the data of A. P. Kropachev et al. [3], carbonate flyschoids in the upper reaches of the Kuranakh River are replaced by an essentially clayey stratum up to 700 m thick, with erratic masses of Upper Devonian and Lower Carboniferous limestones up to 200-300 m in size. In the Burkhalo river basin, one of the authors found unrounded fragments of Upper Ordovician and Silurian dolomites 0.5-0.8 m in size unevenly distributed in homogeneous clayey matrix. The thickness of the stratum containing breccia reaches 250 m; to the north, it is replaced by carbonate flyschoids or is absent. In the Tyry river basin, coarsely fragmental rocks underlay flyschoids and are represented by a 150-m packet of conglomerates with interbeds of limestones. The limestones are bioclastic wackestones and are characteristic of an open-shelf or platform environment (standard microfacies 3, 8, 9). The conglomerates have an erosional lower contact and consist of unsorted limestone clast rounded to various degrees and, more rarely, clasts of dolomites, black flint, and argillites. The sizes of clasts vary from millimeters to no less than 1 m.

The schematic paleogeographical profile for the Early Carboniferous (see Fig. 4) consists of two parts building upon one another along strike (Fig. 1). To the west, sediments of the upper portion of the Tournasian are represented by cross-laminated calcarenites of the Menkyulensk suite formed under lacustrine—alluvial and lagoonal—deltaic conditions, with drift of detritus

from the west [7]. Further to the east, interbeds of shallow water—marine limestones appear; they then disappear and cross-laminated calcarenites with interbeds of conglomerates predominate. Probably, a portion of the detritus built a rise located in the central portion of the profile. The rise lasted for a short interval of time; already by the Visean, limestones of the Muruninsk suite, typical of the open platform, were accumulating here. East of the rise, there are fragments of shallow-water limestones and calcarenites with fauna which are replaced by flyschoid and phthanitic strata; the erratic masses mentioned above appear in the most easterly outcrops. Eight to 10 km to the east of the latter, deposits of the age level under consideration are lacking and Upper Visean strata lie upon Silurian dolomites.

At the Tyry-Burkhala and Sakhara-Allakh-Yun interfluvies, calcarenites of variegated colors and limestones of the western sides of troughs are now destroyed by erosion, but flyschoid strata are replaced toward the east by massive and massive-platey limestones with interbeds of dolomites of calcarenites very similar to deposits of the open platform or the sands at its margin. In this portion of the region, only the thickness of the Lower Visean portion of the flyschoid complex exceeds 1200 m; the role of phthanites is markedly reduced.

Transport of detritus was accomplished primarily by low-density currents: turbidity and liquified flows. The important role of the first of these is confirmed by the widespread occurrence of sediments with Bouma sequences, whereas beds of calcarenites with convolute bedding and flame-type structure are evidence of the existence of the second [15]. Deposits of liquified flow compose lentiform bodies and apparently fill submarine channels. This set of rocks characterizes facies B, C, and D, which are typically found together in the middle portion of a fan [19]. Breccias occurring in the Burkhala river basin are the only (within the region under discussion) documented examples of debris flows south of the Tyry River. Their association with deposits of liquified flow and finely laminated clayey rocks, with subordinate quantities of turbidites (facies A and C predominate over D) are typical of the upper portion of the fans. The topmost portions of the channels are identifiable only in the Tyry river basin, where conglomerate bodies rapidly pinching out along strike cut shallow-water bioclastic limestones. To determine the origin of the erratic masses in the upper regions of Kuranakh River, it is necessary to keep in mind that clasts hundreds of meters in size could hardly have moved as a result of slides or under the influence of debris flows. The erratic masses do not show traces of tectonic working and are only located in direct proximity to the rise situated to the east of them (see Fig. 4) over an area of approximately 6-7 km² [3]. Faults existing at the end of the Tournasian, judging from their relationship with the enclosing sediments, were steeply dipping upthrow faults or dislocations at the time of sedimentation and, apparently, promoted the formation of large scarps. These data most likely indicate a steep, but local avalanche-rockfall around a steep coast sloping 30-60° [15] rather than association of erratic masses with typical olisthostromes around a regional thrust front [3].

The sediments of the Kuranakhsk suite accumulated in several troughs separated by rises, including islands not containing volcanos. The widths of the troughs did not exceed 300 km. Their depths cannot be definitively estimated. Since the widths of the slopes did not exceed 12-15 km (Fig. 4) and liquified flow forms at slopes of no less than 3° [15], the minimal depth can be estimated as approximately 700-800 m. At the same time, the large avalanches on the western slope of the Bagamsk bank are characteristic of scarps with depths from 1 to 4 km. Carbonate-free phthanites abundant in the Lower Visean portion of the Kuranakhsk suite [1] also suggest depths measured in kilometers. However, slopes need to have exceeded 10° for a basin depth of more than 2.5 km; this would have caused formation of numerous grain flows cresting massive beds with flat tops, a predominant orientation of fragments in the direction of the current, and reverse graded bedding around their base [15]. The authors found nothing analogous during their field work and did not find anything similar in descriptions by other investigators. In this connection, it is proposed that the slope could hardly have exceeded 3-5° and, correspondingly, the depth of the basin could hardly have been more than 1.5-2 km.

The sediments of the Lower Carboniferous are distinguished by high sedimentation rates. The sedimentation rate for Lower Visean strata south of the Tyry River, even neglecting compaction during lithification and deformation processes, equals approximately 20-25 cm/1000 years and, when appropriate corrections are introduced [16, 18], equals no less than 35-40 cm/1000 years. Such values are characteristic of avalanche sedimentation and are found on continental slopes, abyssal fans, and river deltas [8, 21]. Such high sedimentation rates are possibly due to the global drop in sea level during the middle of the Early Carboniferous [8, 20, 22]; this should have activated processes of avalanche sedimentation. The presence of a spatially localized system of submarine channels and a regular alternation of facies [19] allows us to assume that the flyschoid-type deposits of the Kuranakhsk suite characterize submarine fans not exhibiting any important differences from submarine fans in terrigenous strata.

* * *

The deposits of carbonate density currents of the Cambrian and Carboniferous have a number of features in common. In both cases, deposition occurred in basins similar to present-day passive continental margins. Nevertheless, the sizes and depths of the depositional basins, like the composition of the sediments, have little in common with modern oceans; moreover, there is a lack of reliable indices of similarity among the latter. Both strata occupy similar positions in formation series. The Cambrian carbonate flyschoids and bitumeniferous calcareoargillaceous rocks of the Inikansk and underlying Akrinsk suites (the lower is Middle Cambrian) lie upon shallow-water sandy-carbonate strata of the Yudomsk series and the Sytyginsk and Pestrotsvetnaya suites (Vendian-Lower Cambrian), and are overlapped by shallow-water pelagic shales of the Atyrbakhsk suite (Lower Ordovician). Flyschoids of the Tournasian-Visean Kuranakhsk suite are underlain by shallow-water limestones of the Kanamytsk suite (lower portion of the Upper Tournasian) and are overlapped by shallow-water pelagic shales of the Senduchensk and Ovlachansk suites (upper portion of the Visean). In both cases, deposits of carbonate density currents thus mark the initial stages in the formation of slopes of deep-water troughs. Rhythmicity of at least four orders can be identified in both the Cambrian and the Carboniferous strata. The thicknesses of the first two rhythms are in the range of millimeters — a few centimeters (laminates) and centimeters to 1.5-2 m (Bouma sequences); these were described above. Bouma sequences are united into rhythm packets with thicknesses of 10-25 m (found only in the Ulakhsk series) and 80-300 m distinguished by specifics of distribution and ratio of clayey to calcareous material. Analogous rhythm packets are classified by Yu. K. Sovetov [9] as first-order flysch structures; he associates their formation with tectonic factors or eustatic fluctuations in sea level. As noted above, the Kuranakhsk suite accumulated precisely during a drop in sea level. During the Middle-Late Cambrian, a rise in sea level took place [8, 20, 22]; tectonic processes and, most of all, intensive downwarping of the bottom within the region of the Ulakhsk series probably played an important role here.

The deposits under discussion exhibit a number of differences as well as similarities. Thus, the Cambrian strata accumulated in a single trough, but the Lower Carboniferous strata accumulated in troughs separated by a series of rises, including islands (one of them is shown in Fig. 4). The Cambrian deposits are associated with a hybrid type of depositional environment (in the terminology of D. Stow et al. [22]), and the Lower Carboniferous deposits are associated with submarine fans. The latter are also distinguished by high sedimentation rates. In conclusion, it is necessary to emphasize that the deposits under consideration are well exposed, widely distributed, and distinguished by a variety of genetic types, including deposits of almost all varieties of density currents (including grain flows). These features make the Cambrian and Lower Carboniferous strata of Sette-Daban a suitable object of investigation for verifying different models of carbonate deposition.

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PATCHAIM TAZHIBAEVNA TAZHIBAEV



Fig. 1. Patchaim Tazhibaevna Tazhibaeva

The geological sciences within the Soviet Union have suffered a heavy loss. On January 6, 1991, after a long illness, Professor Patchaim Tazhibaevna Tazhibaev passed away. She was a major scholar in the field of lithology and weathering crusts, manager of the Lithological Laboratory in the K. I. Satpaev Order of the Red Banner of Labor Institute of Geological Sciences in the Academy of Sciences of the KazSSR, laureate of the Prize of the Council of Ministers of the KazSSR, and Doctor of Geological and Mineralogical Sciences.

P. T. Tazhibaeva was born in 1920 in the mountain village of Kara-Kiya in the Lengerskiy region of the Chimken oblast. After graduating in 1942 from the V. I. Lenin Central Asian State University she studied as an aspirant at the Kazakh Branch of the Academy of Sciences of the USSR. The whole of her subsequent working life was related to the K. I. Satpaev Institute of Geological Sciences, where she worked as a senior research fellow, head of the x-ray laboratory, and later of the lithological laboratory. She defended her candidate's dissertation in 1946, and her doctoral thesis in 1960. In 1967 she was elected as a corresponding member of the Academy of Sciences of the KazSSR. The name of P. T. Tazhibaeva is closely associated with the establishment and development of the lithological sciences in Kazakhstan. Under her guidance and with her personal participation was undertaken the study of Proterozoic, Paleozoic and Mesozoic deposits, weathering crusts, and associated natural resources within Kazakhstan. She had a wide range of scientific interests: lithology, mineralogy, crystallochemistry, paleogeography, and the study of metallic and nonmetallic natural resources.

The editorial board of *Lithology and Mineral Resources*; the Interdepartmental Lithological Committee of the Academy of Sciences of the USSR. Earth Sciences Department of the Academy of Sciences of the Kazakh SSR. Institute of Geological Sciences of the Academy of Sciences of the Kazakh SSR. Translated from *Litologiya i Poleznye Iskopaemye*, No. 5, pp. 142-143, September-October, 1991.

P. T. Tazhibaeva's work on the copper sandstones of Dzhezkazgan, the rhodosite—asbestos-bearing deposits of Kumola, and of nickel-bearing and kaolinized weathering crusts, became well-known in this country. In later years, the Paleozoic hydrocarbon-bearing deposits of the Precaspian basin were studied under her guidance. She was the author of more than 150 scientific works, including four monographs. She possessed several author's prizes. In 1988, P. T. Tazhibaeva was awarded the prize of the Council of Ministers of the KazSSR for her development of a range of physicochemical methods for studying sulfur minerals, and for using the results to develop the national economy.

One of P. T. Tazhibaeva's major contributions was her training of scientific personnel. Twelve aspirants and doctoral students successfully defended their work under her guidance. These included not only geologists, but also physicists working in the field of the structure and kinetics of mineral transformations in sedimentary rocks and weathering crusts.

Patchaim Tazhibaevna regularly represented Soviet lithology at international geological and sedimentological congresses, and symposia on the study of clays. She was the Head of the Kazakhstan Department of the Interdepartmental Lithological Committee for about 30 years, and was a member of the plenum of the Interdepartmental Lithological Committee and the All-Union Commission for the Study of Clays and Clay Minerals. From 1966 she was a permanent member of International Association for the Study of Clays and Clay Minerals, a member of the Japanese Society for the Study of Clays, and also of the Society of Economic Paleontology and Mineralogy (SEPM) of the USA.

P. T. Tazhibaeva was the instigator in Alma Ata of a number of All-Union and republican conferences and symposia, which included participants from foreign countries. She combined her work in scientific organization with her roles as a deputy in the Alma Ata city council, a member of the plenum of the Alma Ata oblast committee of the Communist Party of Kazakhstan, president of the republican committee of the Trade Union for High-School and Scientific-Establishment Workers, Deputy of the Supreme Soviet of the KazSSR, and Deputy President of the Supreme Soviet of the KazSSR.

P. T. Tazhibaeva's merits were highly valued: she was the recipient of two Red Banners of Labor, the Order of Friendship of the People, and with medals and an Honored Citation from the Supreme Soviet of the KazSSR.

A clear memory of this remarkable person, who was a great scholar, communist, patriot, and internationalist will forever remain in our hearts.

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